Bulletin of the Geological Society of America

VOLUME 54 FEBRUARY 1, 1943 NUMBER 2

CONTENTS

	PAGES
Late Paleozoic formations of central Oregon. By C. W. Merriam and S. A. Berthiaume	145–172
Geology and mineralization of the San Antonio mine, Santa Eulalia district, Chihuahua, Mexico. By William Paxton Hewitt	173–204
Stratigraphy and structure of southwest Saypo quadrangle, Montana. By Charles Deiss	205–262
Elasticity of igneous rocks at high temperatures and pressures. By Francis Birch	263–286

Subscription, $10 per year.

Publication Office: Florida Avenue and Eckington Place, Washington, D. C.

Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117th Street, New York, N. Y.

NOTICE.—In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter in the Post-Office at Washington, D. C., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103, Act of October 3, 1917, authorized on July 8, 1918.

PRESS OF JUDD & DETWEILER, INC., WASHINGTON, D. C.

PAPERS IN PRESS FOR FORTHCOMING ISSUES

Earthquakes and Structure in Southern California. By Beno Gutenberg

Seismological Evidence for Roots of Mountains. By Beno Gutenberg

Plagioclase Twinning. By R. C. Emmons and R. W. Gates

Meteorites and an Earth-Model. By Reginald A. Daly

Helium Age Measurements. I. Preliminary Magnetite Index. By Patrick M. Hurley and Clark Goodman

Growth of the North American Ice Sheet During the Wisconsin Age. By Richard Foster Flint

Ice Sheets. By Max Demorest

Alpine Mudflows in Grand Teton National Park, Wyoming. By F. M. Fryxell and Leland Horberg

Studies on Carboniferous Insects from Commentry, France. By F. M. Carpenter

Geologic Interpretation of Gravity Anomalies in the Southern New England-Hudson Valley Region. By Chester R. Longwell

Carboniferous Formations of the Uinta and Northern Wasatch Mountains. By J. Stewart Williams

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 54, PP. 145-172, 1 PL., 1 FIG. FEBRUARY 1, 1943

LATE PALEOZOIC FORMATIONS OF CENTRAL OREGON

BY C. W. MERRIAM AND S. A. BERTHIAUME

CONTENTS

Page

Abstract 146
Introduction 146
Location of area 146
Previous investigation 147
Acknowledgments 148
Stratigraphic section 149
General statement 149
Coffee Creek formation, Lower Carboniferous 149
General description 149
Thickness 151
Stratigraphic relations and age 151
Spotted Ridge formation, Pennsylvanian 151
General description 151
Plant-bearing sandstones and mudstones 152
Conglomerates 153
Marine deposits 153
Magnetite sandstone 153
Cherts 154
Thickness 155
Stratigraphic relations and age 155
Coyote Butte formation, Permian 156
General description 156
Thickness 157
Stratigraphic relations and age 157
Triassic deposits 158
Jurassic deposits of Wade Butte 160
Geologic structure 160
Comparison with other western sections 161
Baker quadrangle, eastern Oregon 161
Western Oregon 162
Redding area, northern California 163
Alaska and British Columbia 163
Central Nevada 164
Geologic history 164
References cited 167
Register of localities 168

ILLUSTRATIONS

Figure Page

1. Index map 147

Plate Facing page

1. Geologic reconnaissance map and sections of Grindstone-Twelvemile Creeks Paleozoic area, central Oregon 172

ABSTRACT

Fifteen miles southeast of Paulina, central Oregon, unaltered Carboniferous and Permian rocks yield much information regarding late Paleozoic history of this region. Three formations are described as follows: (1) the Coffee Creek (Lower Carboniferous) marine limestones and sandstones bearing *Gigantella, Striatifera,* and *Dibunophyllum;* (2) the Spotted Ridge (Pennsylvanian) disconformably above the Coffee Creek, consisting of cherts, sandstones, heavy conglomerates, and mudstones, in part land-laid, yielding a well-preserved flora provisionally assigned to Pottsville; (3) Coyote Butte (Permian) limestones and sandstones with a large fusulinid, coral, and brachiopod fauna, resting unconformably on both the Spotted Ridge and Coffee Creek.

The faunas indicate a closer relationship with northern California, Alaska, and the Old World than with other known areas in North America. Asiatic affinity of the Coyote Butte Permian fauna is particularly strong.

The Paleozoic beds were folded after deposition of the Coyote Butte Permian and before accumulation of the Triassic conglomerates and sands. The Triassic and older strata suffered further compression later. Strikes of folds within the steeply dipping Paleozoics locally depart widely from trends in superjacent Triassic beds.

Abundant coarse sandstone in all formations and heavy conglomerates in the Spotted Ridge Pennsylvanian bespeak high near-by terrain during much of late Paleozoic and conform to structural evidences of diastrophism.

INTRODUCTION

LOCATION OF AREA

The main body of Paleozoic rocks considered lies approximately 30 miles east of the geographical center of Oregon and 15 miles southeast of Paulina, Crook County, in the headwaters of Grindstone and Twelvemile creeks, tributaries of the upper Crooked River. The 20 square miles shown on the accompanying areal map (Pl. 1) lie adjacent to and immediately south of the south-central edge of the Dayville quadrangle. Wade Butte, a portion of which appears on the Dayville quadrangle, is in the northwest corner of the areal map.[1] The region is today one of sparse population and scattered ranches with Suplee 8 miles to the northeast and Paulina 15 miles northwest the nearest settlements. The area of Paleozoic outcrop is largely grassland bordered on the southeast by more elevated timberlands of the Ochoco National Forest. Much of the territory covered by this study lies on the ranch and grazing area of Mr. Orin Mills.

The Paleozoic belt under discussion was formerly covered by a rather thin mantle of Tertiary lavas and tuffs which have since been largely stripped off. Small thin patches still remain, and large pieces of fossil wood from the former volcanic cover have been found on the surfaces of Paleozoic outcrops far removed from the nearest Tertiary beds. It is probable that this small Paleozoic area has long remained high in both structural and topographic sense. As noted by Lupher (1941, p. 223) there does not appear to be evidence that the great thickness of

[1] *See also* map of Ochoco National Forest, U. S. Dept. of Agriculture, Forest Service, 1930, which covers the region of Paleozoic rocks.

Tertiary volcanics and terrestrial beds so characteristic of the central Oregon region ever covered this particular portion of the State. Those volcanics remaining appear to represent only the later Tertiary stage of activity.

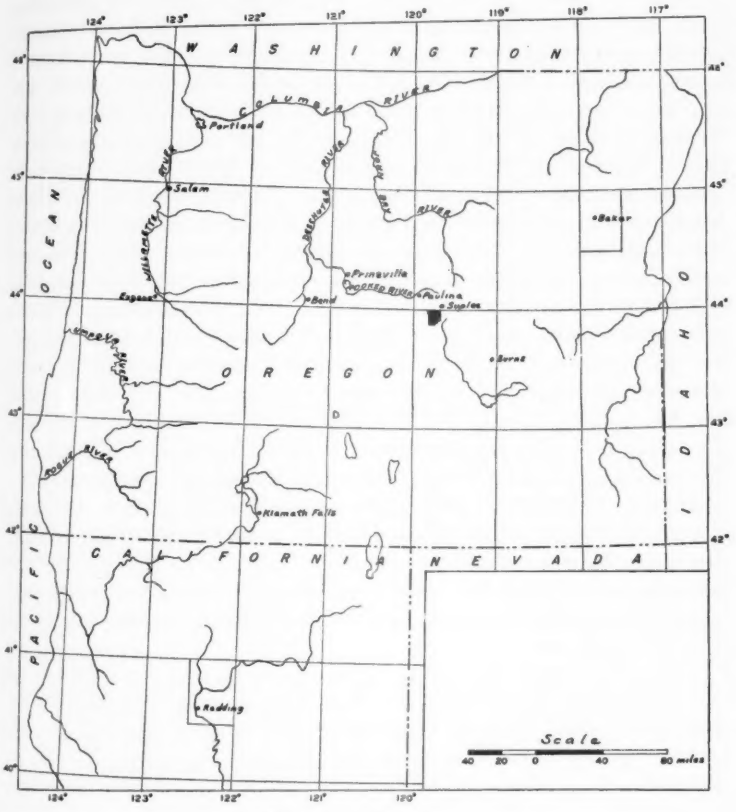


FIGURE 1.—*Index map*
Showing location of mapped Paleozoic area on upper Crooked River southeast of Paulina, Oregon.

In previous papers the region has been referred to as the "Suplee Paleozoic area" and the "Upper Mills Ranch inlier".

PREVIOUS INVESTIGATION

According to Packard (1932, p. 107) the first Paleozoic fossils to be discovered in eastern Oregon were apparently found by Thomas Condon,

whose early contributions to knowledge of Oregon geology are so widely known. Later Washburne (1903) records possible Carboniferous strata on Beaver Creek, Grant County, along the old Prineville-Izee road. Packard (1928; 1932) reviewed the preliminary work which followed his discovery of Lower Carboniferous limestones with *Gigantella* and *Striatifera,* pointing out that Paleozoic strata are found in three areas of the Upper Crooked River basin: (1) the Beaver Creek area about 2 miles northeast of Suplee Post Office, where *Striatifera* is reported, and the limestone masses are overlain by basal Triassic conglomerates with fossiliferous boulders; (2) the Big Flat area about 12 miles east-southeast of Suplee Post Office where nearly vertical limestone beds are overlain unconformably by Triassic conglomerate with limestone boulders bearing fusulinids referred to *Schwagerina;* (3) the largest area southwest of Suplee Post Office in the drainage of Grindstone and Twelvemile creeks, where the major structure was regarded as anticlinal. A relationship of the *Gigantella* fauna found here to the Baird shale fauna of northern California was recognized. While no Permian or Pennsylvanian fossils are listed Packard believed that further work might reveal unconformities and lead to discovery of faunas older or younger than the Lower Carboniferous Baird formation. The presence of Permian fusulinids in Triassic conglomerates resting on the Paleozoics encouraged further search for source rocks of the fossiliferous limestone boulders.

During the summer of 1939 Charles B. Read and Merriam visited the main Paleozoic area to further investigate the flora discovered by Berthiaume and Merriam (Read and Merriam, 1940) in previous years. No additional plant localities were found, but Read collected much well-preserved material; his study of the flora indicates a probable Lower Pennsylvanian age.

Analysis of field evidence by Merriam and Berthiaume (1940) led to the conclusion that at least three formations separated by unconformities are represented in the Paleozoic section; these range in age from Lower Carboniferous to Permian. Terrestrial origin of a part of the middle or Pennsylvanian division was believed probable where plant material is abundant.

ACKNOWLEDGMENTS

Exposures of unaltered Paleozoic strata are rare in Oregon. Hence recognition by Earl L. Packard (1928) of highly fossiliferous late Paleozoic is significant. Following preliminary investigation by University of Oregon parties,[2] Packard invited the writers to conduct a more de-

[2] Master's theses by Norman B. Gonzales (1933) and W. E. McKitrick (1934) at Oregon State College are valuable contributions to the paleontology and geology of the area under consideration.

tailed stratigraphic and paleontologic study. Accordingly several weeks were spent in the field during the summers from 1937 to 1939. The writers thank Dr. Packard for his support in carrying forward the field work and for providing through cooperation of Dr. W. D. Wilkinson a topographic map covering the largest area of Paleozoics. The map prepared by students of the Oregon geology summer camp under direction of Dr. Wilkinson has been of the utmost value. Thanks are due Dr. G. Arthur Cooper of the United States National Museum for his careful study of the brachiopods collected by University of Oregon field parties and by the writers. Investigation of the Carboniferous flora by Charles B. Read of the United States Geological Survey has been of great help in establishing the stratigraphic sequence in this complex area. The writers are furthermore indebted to Dr. L. G. Henbest for identification of fusulinid genera and to Dr. R. L. Lupher for determination of Jurassic fossils from strata above the Paleozoics. During the field season of 1938 valuable assistance was given by Mr. John G. Watson. Field work during 1939 was made possible through a grant made by the Trustee-Faculty Research Committee of Cornell University.

STRATIGRAPHIC SECTION

GENERAL STATEMENT

In the area mapped the deposits range from Lower Carboniferous to Tertiary. The combined thickness of Paleozoics is roughly 3300 feet, while that of the overlying Mesozoic rocks locally exceeds 4000 feet. Tertiary volcanics have largely been removed by erosion, occurring here and there in thin patches or as cap rock of buttes. The stratigraphic relations of formations described are shown in the following section:

Tertiary volcanics
Jurassic sandstones
Triassic sandstones and conglomerates, 3500 to 4000 feet
Deformation (folding)
Permian, Coyote Butte formation; fusulinid limestone, sandy limestones, and sandstone, 900 feet
Deformation (local angular discordance)
Pennsylvanian, Spotted Ridge formation; bedded chert, coarse conglomerate, carbonaceous sandstone, and carbonaceous mudstone; in part terrestrial with Lower Pennsylvanian flora, 1000 to 1500 feet
Disconformity
Lower Carboniferous Coffee Creek formation; marine limestones and calcareous sandstones with *Gigantella, Striatifera,* and *Dibunophyllum,* 900 to 1000 feet

COFFEE CREEK FORMATION, LOWER CARBONIFEROUS

General description.—The Coffee Creek formation is named for exposures on a minor tributary of this name which enters Grindstone Creek south of Wade Butte. At the type section in sec. 30, T. 18 S., R. 25 E. ¼ mile east of the spring at Mills sheep camp a line of limestone outcrops

trend roughly northeast-southwest. In general the formation consists of well-bedded fairly pure limestones, carbonaceous limestones, argillaceous to sandy limestones, and calcareous sandstone. Exposures of the type Coffee Creek area were traced intermittently along the strike for a distance of about 1¼ miles from locality 93 on the south to locality 98 on the north.

A conspicuous linear outcrop at the type section (loc. 2) shows about 75 feet of strata dipping about 53° NW. Within the upper 40 feet very dark gray to black argillaceous and carbonaceous limestones are in places rather thin-bedded. These deposits are locally packed with the brachiopod *Striatifera*. Forty-five feet from the top of the limestone section is a bed of very sandy limestone with *Striatifera* immediately below which lies the main *Gigantella* horizon. The unusually large productid *Gigantella* is very profuse here in a fine-grained and rather pure limestone of deep neutral gray; the large brachiopod is apparently restricted at this point to a bed varying in thickness from one to two feet. Within and below the *Gigantella* bed solitary and compound rugose corals are abundant; these are embedded in the gray limestone or immediately below in argillaceous limestones weathering to a light grayish brown. In the lower 30 feet of this section these argillaceous layers are interbedded with the purer gray limestones, while toward the base sandy limestones and calcareous sandstone predominate. Throughout the entire 75 feet *Striatifera* is the most common fossil, occurring in the lower sandy beds as well as the higher strata. The prominent limestone outcrops of the type section probably represent the higher portion of the formation.

In general, exposures of the Coffee Creek are not numerous in view of the subdued and grassy character of much terrain underlain by this formation. Most of the better outcrops are in the resistant limestones near the *Gigantella* horizon which stand above grass-covered slopes formed by sandy facies of this division. Nearly all known belts of Coffee Creek appear to represent axial portions of rather tight anticlinal folds variously dissected. While the best and most fossiliferous outcrops of Coffee Creek are limestones of its upper part, probably well over half the entire formation is calcareous sandstone, but little of which is exposed in unweathered condition. Pure limestones have not been found in the middle and lower portions of the formation; deposits of these divisions where visible at the surface include greenish pebbly sandstones and sandy limestones with *Striatifera*. Such strata are well shown at locality 3. Patchy distribution of the purer limestone outcrops in the upper part of the formation suggest that they represent more or less extended lenses within arenaceous and argillaceous limestone facies. Ordinarily a single limestone outcrop cannot be followed continuously

along the strike for more than ¼ mile, but the same type of limestone representing approximately the same faunal horizon may appear again in line of strike a few hundred feet beyond. That some such cases are structural and not due to lensing out can hardly be doubted.

Thickness.—Estimates of thickness are greatly handicapped by deformation, and nowhere has the base of this formation been recognized. Conservative figures of 900 to 1000 feet are based on width of outcrop in the anticlines on Coffee Creek, and north of Coyote Butte, where in the last instance the strata stand in nearly vertical position.

Stratigraphic relations and age.—The Coffee Creek formation represents the oldest Paleozoic division recognized in this region. Judging from the lithologic and paleontologic criteria its relationship to the overlying Spotted Ridge Pennsylvanian is one of disconformity. Overlap of the partly land-laid Spotted Ridge upon the Coffee Creek is suggested by distribution of the two units, though stratigraphic evidence is inadequate. In some localities the Coffee Creek is directly overlain by either Permian or Triassic beds. These unconformities are discussed below. The age of the *Gigantella* horizon is based on the listed fauna and is Lower Carboniferous, roughly Viséan, in terms of the British succession. That the lower sands of this division are not older than Lower Carboniferous is attested by presence of the productid *Striatifera*, not known in strata of greater age. The preliminary faunal list is as follows:

Dibunophyllum oregonensis Merriam
Lithostrotion (Lithostrotion) packardi Merriam
Lithostrotion (Siphonodendron) oregonensis Merriam
Campophyllum readi Merriam
Gigantella sp.
Striatifera sp.
Spirifer cf. *striatus* (Martin)
Tetrataxis sp.
Small loxonemoid gastropods
Lithistid sponge spicules

All but the spirifer are very abundant, *Striatifera* the commonest form ranging throughout. *Gigantella* has been found only in the key limestone bed about 45 feet below the top of the formation as shown at locality 2, where the bed is crowded with these shells in association with the corals listed. The Foraminifera, small gastropods, and sponge spicules are abundant at the same horizon; they are silicified and can be prepared by the acid-etching treatment.

SPOTTED RIDGE FORMATION, PENNSYLVANIAN

General description.—Sediments of the Spotted Ridge formation are exceedingly variable in both vertical and horizontal direction, ranging

from compact mudstones and cross-bedded sandstones to very coarse boulder conglomerates. Locally much bedded chert is present. The members thicken and thin in various directions or may pinch out completely within a short distance. Good exposures are rare because the outcrop belts are in alluviated valleys or somewhat depressed areas between strike ridges formed by more resistant Permian rocks. The best exposures are in the type section on the west flank of Spotted Ridge, extending south about 2 miles to locality 83.

Plant-bearing sandstones and mudstones.—Sandstones with much carbonaceous material and recognizable plant remains appear to range throughout the formation, while the lenticular plant-bearing mudstones were found only in the upper part. The lowest sands with plants occur at locality 96 near the Coffee Creek contact. These basal deposits are sands of medium-grain and light neutral-gray color on fresh surface, weathering to various tones of limonite brown. Feldspar grains are abundant. Plant remains range from finely divided carbonaceous debris to flattened stems several inches in length. That a similar facies persists locally or recurs in some situations is indicated by the presence of plant-bearing beds of much the same character at locality 108 within about 75 feet of the top of this division, possibly 900 feet stratigraphically above the locality first mentioned.

Lenticular distribution of sedimentary types is apparent within the plant-bearing deposits. At plant locality 115 on the west side of Spotted Ridge are fine sandstones, siltstones, and mudstones in varying order within a thickness of 5 feet. Intergradation from one layer to another is recognized. Following the same approximate horizon southward a distance of ½ mile to locality 7 one finds cross-bedded sandstone lenses within a heavy conglomerate including 2-foot boulders. Some of the finer sands here contain plant remains in direct association with small nuculid bivalves, small gastropods, and what appears to be a scaphopod about ½ inch in length.

The lenticular mudstones and siltstones at localiy 115 are of an unusual shade of medium-grayish olive green. Lamination is usually not well defined, and fissility is undeveloped. The sediment is very compact and brittle. In general the plants lie more or less parallel to bedding though occasional leaves and stems are decidedly oblique in position. The plants occur as coaly films; stems are much flattened but retain evidences of vascular structure. Impressions of leaves below the carbonaceous films are sharply defined. The fine state of preservation of the dark leaves surrounded by a relatively light matrix lends added value to the plants for study purposes. While certain tongues or lenses within

the plant-bearing beds are marine or brackish it is believed that most of the sediments in this facies are land-laid, though an estuarine origin in part is not unlikely. The plants do not appear to have suffered transportation; in fact some calamite stalks with whorls of twigs appear to be essentially in position of growth.

Conglomerates.—Coarse unsorted conglomerates appear here and there throughout most of the Spotted Ridge section, associated both with the plant-bearing deposits and the marine facies. The fragments are well rounded and range from pebble size to large boulders with a diameter of at least 2 feet. Conspicuous among rocks composing the boulders are intrusives of dioritic character with andesitic and dacitic types subordinate.

Marine deposits.—Marine clastics are particularly well developed within the middle portion of the formation, being well shown at localities 5 and 54. Coarse pebbly sandstones and pebble conglomerates are conspicuous. The pebbly sandstones are composed in large part of well-rounded light-green and light-grayish chert fragments including also a high percentage of feldspar grains. Larger pebbles have a diameter of ¼ inch. Calcite is present as cementing material and as fragments of invertebrate organisms, particularly crinoids. Quartz appears to be relatively rare. Remains of crinoids, gastropods, and undetermined brachiopods indicate a marine environment. An internal mold of an indeterminate turriculate gastropod 1½ inches long was found. The large amount of crinoidal debris in the clastic environment is surprising. Columnal segments with a diameter of ⅝ inch are very common as sharply defined molds. Coarse conglomerates with boulders of basic intrusives are closely associated with the marine sandstones and pebbly deposits at several localities. These are well shown at localities 5, 6, and 10.

Magnetite sandstone.—Two horizons of unusual black magnetite sandstone have been recognized in the Spotted Ridge formation. The lower zone at locality 11 north of Grindstone Creek shows the greatest concentration of magnetite grains in a ½-inch bed within coarse pebbly cross-bedded sands. The grains of the magnetite layer itself are partly rounded and of medium size. Magnetite grains make up about 80 per cent of this sand with feldspars comprising much of the remaining 20 per cent.

The higher magnetite sands lie within 50 feet of the top of the formation near locality 7. The lateral extent of this lens is not known, but at one point the main magnetite bed is about a foot thick. In this bed subrounded to angular magnetite grains constitute 80 per cent of the

sediment. These sands also are medium-grained and contain carbonized plant remains.

Cherts.—Chert or silicified clastic sediments are widespread. The greatest development is in the north-south ridge which continues southward to Grindstone Creek in sections 27 and 28, T. 18 S., R. 25 E. in the northeastern portion of the area. This belt of folded siliceous deposits is nearly a mile wide in places. The cherts show well-defined and rather thin bedding. They are intensely folded and fractured, showing many local minor flexures such as are common in areas of highly disturbed bedded cherts or incompetent siliceous shales. The deposits often have a light green or grayish green color. Toward the north end of the area where this belt of chert enters the Dayville quadrangle the siliceous deposits are closely associated with rather coarse sandy clastics and lie adjacent to outcrops of a plutonic intrusive. Similarly on Grindstone Creek the large plutonic mass northwest of Mills Ranch is close to the broad cherty belt. A somewhat less extensive chert outcrop south of Twelvemile Creek extends from the vicinity of locality 58 southward and westward along the north slope of Coyote Butte. While the distribution of these cherts indicates that they lie chiefly within the limits of the Spotted Ridge formation, it is unlikely that they represent a single stratigraphic zone. On the north flank of Coyote Butte the cherts are apparently several hundred feet thick, lying between the plant-bearing beds of locality 38 and the base of the fusulinid limestones regarded as the lower member of the Coyote Butte Permian. At Spotted Ridge (locality 115) in the northern portion of the area no cherts were recognized in this stratigraphic interval. The greatest development of chert appears thus to be localized. The possibility that it may be secondary and related to igneous activity as a product of secondary silicification has been entertained. However, the distinctly bedded character of the deposit at several localities would seemingly favor an origin as primary chert, though perhaps in such case still to be correlated with local contemporaneous igneous activity. Highly siliceous and cherty deposits are found in the Coyote Butte Permian and in the overlying Triassic. The cherty limestones of the Coyote Butte formation occur north and south of Twelvemile Creek near the west border of the Paleozoic area. Here, as at locality 13, black chert is irregularly interbedded with the limestone. The siliceous deposits within the Triassic are well shown at locality 21 southeast of Wade Butte. These small localized areas within the Triassic clastics probably represent areas of silicification related to igneous stocks.

Thickness.—In view of intense folding and lack of continuous outcrop it was not found possible to arrive at accurate figures regarding thickness. Variability in lithology from place to place along the outcrop belts suggests great differences in thickness of the entire section and of its individual members. This applies particularly to the conglomerates, and to the cherts, which last are entirely absent at certain localities in horizons where they would be expected. Judging roughly from width of outcrop the formation at the type section, extending through locality 115, is probably 1000 feet thick. Southward along Twelvemile Creek the thickness is at least 1000 feet, while on the north slope of Coyote Butte it may, with the included cherts, exceed 1500 feet.

Stratigraphic relations and age.—The lowest exposed Spotted Ridge beds recognized lie in the vicinity of locality 96 where plant-bearing sands were found within a few feet of a well-exposed limestone bed of the *Gigantella* horizon and therefore near the top of the Coffee Creek Lower Carboniferous. Uplift and emergence following deposition of the purely marine upper Coffee Creek is suggested by the probable terrestrial or estuarine nature of these lower Spotted Ridge deposits. Nothing conclusive has, however, been determined regarding the magnitude of the break separating Coffee Creek from Spotted Ridge beyond the fact that the marine faunas of the upper Coffee Creek are of Lower Carboniferous age, while floral evidence suggests a Lower Pennsylvanian age for the upper Spotted Ridge beds.

The Coyote Butte Permian is unconformable upon the Spotted Ridge. Judging from fossil evidence there is in all probability a hiatus representing a portion of the Lower Permian and perhaps most of the Upper Pennsylvanian. A sharply defined contact between Spotted Ridge sandstones and relatively pure limestone of the Coyote Butte is exposed at locality 105 in the southwestern portion of the area. There is here a slight angular discordance between the two divisions. The Coyote Butte formation is regarded as marine in contrast to the partly estuarine, terrestrial and alluvial or delta-plain character of the underlying Spotted Ridge.

While no determinable marine fossils were discovered in the Spotted Ridge formation, reliable evidence regarding its age is provided by the flora found at locality 115 in a horizon within the upper 400 feet. According to Read and Merriam (1940, p. 109)

"Most abundant and striking elements in the flora are several species of *Sphenopteris* and *Dactylotheca*. These, in fact, constitute the entire fern population. Representatives of genera which commonly have high stratigraphic value (e.g., *Neuropteris, Mariopteris, Pecopteris*) are unknown and in consequence make more difficult an age pronouncement. Associated with the matted fern fronds are common *Calamites* stems, *Asterophyllites*, and the stems, leaves, and strobili of a form

temporarily assigned to *Phyllotheca* because of its obvious affinities with *P. ralli* Zeiller from the Westphalian of Heraclea in Asia Minor. Conspicuous although not extremely abundant is a species belonging to the very rare gymnospermous genus *Dicranophyllum*. Material referable to two new genera of cordaitalean affinities also occurs, as do some problematica not definitely assignable at present."

Read and Merriam conclude that the flora is Lower Pennsylvanian, stating, however, that the possibility of its being Upper Pennsylvanian cannot be completely ruled out. If the flora ultimately proves to be Lower Pennsylvanian, it is likely that no great gap exists between the Coffee Creek and Spotted Ridge formations.

COYOTE BUTTE FORMATION, PERMIAN

General description.—The youngest Paleozoic beds of the area comprise a sequence in which massive limestones form the most conspicuous exposures. These produce prominent ridges, buttes, and small circular hills or knobs subsidiary to the main ridge slopes. Steeply dipping strata forming the crest of Coyote Butte near the southern limit of the map constitute the type section of the division. The Coyote Butte beds here lie in the north limb of a tight syncline overturned toward the south. Another bold outcrop of the formation is found in the belt of limestone extending north-northeast about 4 miles from the vicinity of Tuckers Butte to the spring at locality 92. A third area of Permian beds includes the limestone exposures in the northeast portion of the map, one tongue of which extends southwest beyond Twelvemile Creek where it is covered by Tertiary lava.

At the type section the lower portion of the Coyote Butte is generally a light olive-gray limestone, often crinoidal and locally containing abundant fusulinids. Higher in the section, at the summit of Coyote Butte, the limestone becomes purer, finer-grained, deep olive-gray and more distinctly bedded. In this upper portion there are fewer fusulinids, while brachiopods are common.

The Coyote Butte limestone throughout the area is essentially similar to that of the type locality. Coarse crinoidal debris is abundant at several localities not far above the lower contact of the formation. Conglomeratic phases of the basal limestone also occur locally. At locality 73 in the eastern part of the area rounded chert pebbles and subangular chert grains are distributed throughout a crinoidal and fusulinid-bearing limestone. In another conglomeratic phase at the base of the limestone near locality 80 on the west side of Spotted Ridge green, white, red, and black chert pebbles become so abundant as to form a calcareous conglomerate. These pebbly facies probably represent a local reworking of underlying Pennsylvanian sediments.

Limestones of the Coyote Butte formation are interbedded with large amounts of sandstones. Furthermore the discontinuous nature of some of the limestone exposures leads to the impression that these deposits are lenticular within the more arenaceous facies. Generally speaking, exposures of the sands are poor, while the limestones form prominent linear or circumscribed exposures. Northeast of locality 123 in the center of the map several zones of sandstone float alternate with fusulinid limestones.

Large amounts of insoluble materials, such as chert and sand grains, occur locally in the Permian limestone. At locality 68 in the northeastern portion of the area dark, irregular chert layers up to 10 inches thick are interbedded with the limestone. In other occurrences the siliceous material replaces the limestone as irregular patches and permeates it in varying degree. In some instances, as along the southwest side of Spotted Ridge, the limestones are dolomitic and in extreme cases are apparently recrystallized, giving the rock a light-gray or pale-cream color.

Thickness.—At the type section on Coyote Butte approximately 900 feet of the Permian beds is exposed. Estimates of thickness on Spotted Ridge and north of Tuckers Butte are approximately the same. However, exact thicknesses cannot be given since bedding is poor and folds within the massive limestone are difficult to work out. Where the formation is largely arenaceous the exposures are poor. Unconformable relation of the Permian to overlying beds further eliminates true thickness estimation.

Stratigraphic relations and age.—The Coyote Butte formation is unconformable on the Spotted Ridge Pennsylvanian, as suggested by pinching out of the Pennsylvanian strata in sec. 5, T. 19 S., R. 25 E., where the contact between Coyote Butte limestone and the Lower Carboniferous Coffee Creek formation is apparently depositional. At locality 105 north of Tuckers Butte an exposure of the lower contact of the Permian shows a slight angular discordance and a definite truncation of the conglomeratic and sandy beds of the underlying formation. On the basis of lithology and position the latter beds are presumed to be the Spotted Ridge formation. Furthermore, on the west side of Spotted Ridge and at several other localities the basal Coyote Butte strata are very pebbly limestones and calcareous conglomerates, probably indicating reworking of subjacent Pennsylvanian clastics.

Fusulinids, corals, and brachiopods from the Coyote Butte formation indicate a Permian age for these beds. Several faunal zones are undoubtedly represented but have not been differentiated in view of the

complex structure and lenticularity of sedimentation. More refined studies of structure, stratigraphy, and sedimentation will be required to work out the details of this zonation. Since the Spotted Ridge plant-bearing beds a short distance below the Coyote Butte formation are regarded as Lower Pennsylvanian, a hiatus of some magnitude is indicated between the two formations. Nearly all the localities in the lower part of the formation have yielded a new species of *Parafusulina*. In addition, species of *Schwagerina* are found at several localities as well as forms tentatively referred to *Fusulinella* and *Triticites*. Field evidence shows that several of the fusulinid types are either associated or occur in almost the same horizon. The fusulines imply that the Coyote Butte is not lowest Permian.

Dr. G. Arthur Cooper is now completing a detailed study of the Coyote Butte brachiopods and reports that most of the species are of Asiatic affinity; a few are almost exact identities with Russian forms from the Urals and Timan. According to Cooper the brachiopods indicate Lower Permian. The following identifications were supplied by Cooper:

Productus cf. *P. mammatus* Keyserling
P. aff. *P. porrectus* Kutorga
Avonia tuberculata Schellwien
Linoproductus cf. *L. sinuata* King
Juresania aff. *J. juresanensis* Tschernyschew
Waagenoconcha n. sp.
Krotovia pustulata Keyserling
Keyserlingina sp.
Marginifera cf. *M. involuta* Tschernyschew
Rhynchopora n. sp.
Camarophoria mutabilis n. var.
C. biplicata Stuckenberg
C. karpinskyi Tschernyschew
Nototthyris nucleola Kutorga
Nototthyris n. sp.
Martiniopsis sp.
Spiriferella n. sp.

In a recent study of rugose corals from the Coyote Butte formation Merriam (1942) has described the following forms:

Waagenophyllum washburni Merriam
Waagenophyllum ochocoensis Merriam
Waagenophyllum sp. a.
Waagenophyllum sp. b
Lithostrotion (Lithostrotionella) occidentalis Merriam
Lithostrotion (Lithostrotionella) berthiaumi Merriam

The species of *Waagenophyllum* support a Permian age and point to an Old World relationship of the faunas.

The ammonoid *Eoasianites merriami* Miller and Furnish was found in the region covered by the present survey, but whether it came from the Spotted Ridge Pennsylvanian or the Coyote Butte is not known. According to Miller and Furnish (1941, p. 541-542) it appears to be of either Upper Pennsylvanian or Lower Permian age.

TRIASSIC DEPOSITS

A broad outcrop of conglomerates and sandstones underlies the Jurassic strata of Wade Butte and overlies the Coffee Creek Lower Carboniferous

1½ miles to the southeast. Rocks of similar character and stratigraphic position are found on the east side of Spotted Ridge and in a synclinal belt farther south between Grindstone and Twelvemile creeks. A third area of these coarse clastics is found along the eastern margin of the Paleozoics on Grindstone Creek. Conglomerates, sandstones, and cherty beds apparently overlying normally the Permian south of Coyote Butte may possibly belong to the sequence just mentioned, but the isolated occurrence and lack of good lithologic or fossil evidence leaves their relationship somewhat in doubt. As yet there is no fossil evidence in the area mapped which would definitely establish a Triassic age for any of these conglomeratic beds. However, at Wade Butte they underlie Jurassic strata of the Colpitts group (Lupher, 1941, p. 227). Here, as in other localities, they appear to correspond to Triassic deposits recognized by Edward Schenk (*see* Packard, 1932, p. 109) which usually contain large limestone boulders with Permian fusulinids. The possibility that some of the upper part of this supposedly Triassic sequence is pre-Colpitts Jurassic must, however, be entertained, but the beds are here referred to as Triassic.

The area of Triassic rocks southeast of Wade Butte shows a series of massive, fine to coarse conglomerate beds, and gray or greenish sands and pebbly sands which usually weather brownish. The conglomerates form the major portion of the section and seem to be lithologically distinct from the Pennsylvanian conglomerates in the presence of limestone cobbles. Large igneous boulders found in the Spotted Ridge conglomerates seem to be absent from the Triassic sequence. Very angular green and black chert fragments are the chief constituents of much of the conglomerate with occasional pebbles of white or gray chalcedony, clear quartz, fine-grained sandstone, and aphanitic igneous rock. Some of the chert pebbles are much fractured and veined. The sandstones at locality 21 are gray, slightly calcareous, and contain angular chert and quartz grains, abundant and rather fresh orthoclase, and a few grains of a cryptocrystalline igneous rock.

The thickness of the Triassic sediments is estimated from the monoclinal sequence southeast of Wade Butte to be about 4000 feet. This includes the conglomeratic beds unconformably overlying the Coffee Creek formation and dipping northwest beneath the lowest fossiliferous Jurassic beds near the base of Wade Butte. In the synclinal area between Grindstone and Twelvemile creeks, a much thinner sequence is preserved, possibly about 2000 feet.

A prominent angular unconformity exists between the Paleozoic beds and the Triassic conglomerates, although at no place was the actual contact observed. Southeast of Wade Butte and north of Twelvemile Creek,

mapping shows that the Triassic directly overlies the Coffee Creek formation, while east of Spotted Ridge and to the south of Grindstone Creek the conglomerates rest on the Coyote Butte beds. Overlap is further indicated by small isolated residuals of Triassic conglomerate resting on various older horizons (Locs. 4, 94).

Evidence of unconformable relation to the Paleozoic strata is afforded by the presence of large limestone boulders in the Triassic conglomerate. Many of the boulders contain brachiopods and fusulinids characteristic of the Coyote Butte strata, while others in association with the first bear *Striatifera* and corals from the Coffee Creek formation. In certain limestone cobbles a fusulinid of slightly older aspect than those found in place within the lower Coyote Butte suggests former existence of a late Pennsylvanian or early Permian horizon in this general region. Its absence from the stratigraphic sequence in the local area examined implies removal by pre-Coyote Butte erosion or nondeposition of the strata in the Grindstone-Twelvemile creeks area.

JURASSIC DEPOSITS OF WADE BUTTE

On the south and east sides of Wade Butte several hundred feet of Jurassic rocks rest upon the Triassic conglomerate sequence. Toward the base of this section the deposits are uniformly fine grained grayish sandstones which weather to grayish limonite brown. They are highly calcareous, locally approaching arenaceous limestone. In a characteristic specimen cloudy calcite makes up from 50 to 60 per cent, while the remainder consists of rounded grains of chert and quartz. Lithologically the Jurassic sandstones are readily distinguishable from those of the presumed Triassic and from sands of the underlying Paleozoic in their finer grain and absence in this area of a greenish cast. Fossils are abundant near the bottom of the section at Wade Butte. Dr. R. L. Lupher has determined these lower fossils as pertaining to the Weberg formation of the Colpitts group (Lupher, 1941, p. 227) or early Middle Jurassic. As the Colpitts group in central Oregon is underlain by older Jurassic sediments, it is not improbable that higher horizons of the conglomeratic section here called Triassic may actually be Jurassic.

GEOLOGIC STRUCTURE

Indications of pre-Permian unconformities involving slight folding and emergence have been mentioned in discussion of the stratigraphy. However, the large structural features appear to be the result of at least two major periods of deformation recognizable within this area. The first occurred after deposition of the Coyote Butte formation and prior to accumulation of the Triassic conglomerate. Compression during this

earlier deformation resulted in a series of folds striking northeast and north-northeast. Since dips in the Paleozoic beds are generally 60° or greater and the folds are rather closely spaced, this first compression must have been quite intense.

Structures formed during the second recognizable period of major folding were produced by movements affecting the Triassic deposits and possibly the Jurassic as well. Certain folds produced during this interval have a more northwesterly strike and cut across the previously formed flexures. This is well shown by the syncline in beds of supposedly Triassic age between Grindstone and Twelvemile creeks. The axis of this syncline strikes for some distance about N. 30° W., although farther north the strike becomes more nearly north. The same structure appears to have been influenced by the earlier trends or to have been subjected to stress from a different direction. In this latter area near the north end of Spotted Ridge the Paleozoic beds and possibly the Triassic are overturned toward the west. Although the dips in many exposures of the Triassic beds are greater than 40°, the folds seem to be more open and less steep than those in the Paleozoics. It is evident that this second deformation took place before the region was covered with the Tertiary lavas, since the latter do not show steep dips and rest unconformably on the folded Mesozoic rocks at Wade Butte.

That the strata of this general region suffered additional compression during an interval between Jurassic and Triassic seems likely in view of Lupher's (1941, p. 228) recognition of wide divergence between the axes of folds in Triassic and Jurassic rocks.

Presumably some faulting accompanied major periods of folding, but no actual evidence of such displacement was observed. Slickensided surfaces and brecciation can be seen at numerous exposures, but mapping fails to support evidence for major faulting. Poor bedding, strong folding, and lack of mappable subdivisions within the formations help to obscure such structural features.

In the Baker quadrangle, approximately 100 miles northeast of the Grindstone Creek area, Gilluly found Paleozoic rocks metamorphosed by pre-Tertiary dynamic forces producing westerly strikes. This deformation in eastern Oregon may be related to folding in central Oregon, but as yet little evidence is available for dating it in terms of the intervals of folding in the latter region.

COMPARISON WITH OTHER WESTERN SECTIONS

BAKER QUADRANGLE, EASTERN OREGON

In the Baker area 100 miles northeast of Grindstone Creek three Paleozoic formations have been described by Gilluly (1937). The Burnt

River schist of indefinite age shows little similarity to any of the formations under discussion beyond the occurrence of limestones and conglomerates. Presence of much altered tuffaceous material is distinctive. Absence of fossils, intense metamorphism, and lack of evidence concerning stratigraphic position make further comparison fruitless.

The Elkhorn Ridge argillite contains "*Fusulina*" and is apparently at least in part Pennsylvanian. It is perhaps comparable in age to the Spotted Ridge formation of central Oregon. Having suffered alteration and more intense deformation the Elkhorn Ridge offers no close lithologic similarity to the beds on Grindstone Creek. In both cases, however, clastics predominate. In contrast to the Spotted Ridge formation, conglomerates are not conspicuous in the Elkhorn Ridge and apparently do not contain heavy boulders of igneous rocks. In the Baker area presence of limestone as well as bedded chert and argillite suggests a marine origin for much of the Elkhorn Ridge argillite, while in the Pennsylvanian of central Oregon there is strong evidence for terrestrial and estuarine deposition of a large portion of the Spotted Ridge formation. Nevertheless, a considerable amount of chert near the top of the Pennsylvanian sequence at Coyote Butte recalls the large amount of chert and cherty argillite in the Elkhorn Ridge. While no evidence for contemporaneous volcanic activity has been found in central Oregon, the Elkhorn Ridge contains a large proportion of pyroclastic material.

Large quantities of volcanic material in the Clover Creek greenstone distinguish it from the Permian Coyote Butte. The large percentage of limestone in central Oregon is in direct contrast to its small proportion in the Baker quadrangle. According to Gilluly the lens form of limestone bodies in the latter area is presumably due to later shearing and squeezing, not to original deposition. In the less disturbed Coyote Butte formation some of the circumscribed limestone bodies are probably true depositional lenses. No detailed faunal comparison is yet possible as many of the available identifications of Clover Creek forms are generic only. According to Girty (Gilluly, 1937, p. 26-27) the later fauna "appears to be more nearly allied to the Permian faunas of Alaska than to any that have come under my observation from the United States".

WESTERN OREGON

Certain limestone belts in the region of Grants Pass, Oregon, have been considered by Diller (1909; 1914) to be Devonian and Carboniferous, although little conclusive evidence as to their exact age is available. Mentioned as associated with these limestone lentils are slates and tuff beds.

REDDING AREA, NORTHERN CALIFORNIA

Faunally the Coffee Creek formation of central Oregon is related to the upper part of the Baird formation of northern California in the presence of the distinctive brachiopod genus *Gigantella* (Smith, 1894a; 1894b). However, the Oregon exposures give no indication of volcanic activity comparable to that in the vicinity of Redding during deposition of the Baird formation.

In both areas there is evidence of marine withdrawal during upper Carboniferous time. At Redding the regression is represented by an interval of nondeposition between the Lower Carboniferous Baird formation and the Lower Permian McCloud limestone. In central Oregon, the accumulation of the coarse Pennsylvanian clastics and plant-bearing strata also points to predominantly though not entirely nonmarine conditions.

The northern California Permian sequence (Wheeler, 1935) includes strata containing *Pseudoschwagerina* and may, therefore, in part be older than any Permian strata exposed in the central Oregon series. The lowest strata of the Coyote Butte formation would appear on the basis of fusulinids alone to be not older than middle McCloud, which contains *Parafusulina*. Both formations are characterized by massive accumulations of calcareous material. Although in central Oregon a fusuline occurring in limestone boulders of the Triassic conglomerate suggests the presence of Permian older than the *Parafusulina* zone, no beds with a *Pseudoschwagerina* fauna were found.

It appears likely that the Nosoni formation of the Redding region belongs to a slightly younger Permian zone than the highest beds at Coyote Butte in central Oregon. Again, the Nosoni formation differs in containing a great quantity of volcanic debris and very little calcareous material. Further, it is reported to rest uneonformably on the McCloud limestone.

ALASKA AND BRITISH COLUMBIA

Portions of the widespread Mississippian deposits of Alaska, such as the Lisburne limestone, are probably contemporaneous with the Coffee Creek beds of Oregon. In southeastern Alaska (Buddington and Chapin, 1929) the Lower Carboniferous has yielded *Gigantella*. More precise correlation, however, must await study and comparison of the faunas in question.

Further similarity between the Oregon sequence and that in Alaska lies in the possibility that the Pennsylvanian deposits of both regions are largely nonmarine. According to Moffit (1938, p. 28) G. H. Girty [3]

[3] *See also* Williams, J. S. (1941).

had concluded "that marine sediments of Pennsylvanian age are almost unrepresented in Alaska and reassigned faunas once thought to be Pennsylvanian either to the Upper Mississippian (the widely distributed Lisburne fauna) or to the Permian".

The plant-bearing deposits of the Nation River formation in the Yukon Valley occupy a stratigraphic position comparable to that of the Spotted Ridge formation in Oregon.

The Permian Cache Creek series of British Columbia (Crockford and Warren, 1935) contains *Neoschwagerina* and an undescribed *Codonofusiella* together with ammonoids which have been regarded as probable equivalents of forms in the Word formation of Texas. The association is therefore apparently younger than the Coyote Butte assemblage in central Oregon, but, since the fossiliferous beds in the Cache Creek series are near the upper part of the sequence, equivalents of the Oregon Permian may yet be found in lower beds. Possible equivalence of the Cache Creek fauna to that in the Nosoni and Robinson formations of northern California would suggest that similar faunas might occur in the Oregon sections if not eroded before Mesozoic deposition.[4] Apparently most of the formations of this age tend to contain considerable volcanic material. In various parts of Alaska, similar Permian beds are widespread, but close correlation is not yet attainable.

CENTRAL NEVADA

The Carboniferous section near Eureka, Nevada, not only lacks close faunal similarities to the Oregon section but also indicates the chief strictly marine deposition to be in the Upper Carboniferous rather than in the lower portion as is common in known Pacific Coast sections. The Permian of central Oregon and central Nevada may prove to have similar faunal assemblages, but further study is necessary before definite conclusions may be reached.

GEOLOGIC HISTORY

Deposits of the Coffee Creek formation indicate that during the upper part of Lower Carboniferous time the central Oregon region was occupied by a shallow sea which, judging from foreign character of corals and brachiopods, had immediate connection with the Old World, probably through Alaska. Lime mud conditions in later Coffee Creek time were particularly favorable for vigorous growth of rugose corals and very large heavy-shelled brachiopods. A great deal of the sediment is, however,

[4] A species of the genus *Yabeina* has recently been reported by Thompson and Wheeler (1941) from a cobble in Tertiary gravels near Madras, north-central Oregon. Two species of *Yabeina* are reported also from the upper Cache Creek series.

arenaceous and less fossiliferous. Evidently the sandy deposits were laid down at no great distance from shore, and adjacent lands stood sufficiently high to yield large quantities of rather coarse arenaceous detritus. A great abundance of siliceous sponge spicules in the purer limestone facies with *Gigantella* suggests either temporary deepening of the waters or base leveling of lands at a time when sands were not reaching sites of lime mud deposition.

Emergence followed not long after accumulation of the *Gigantella* limestones for plant-bearing clastics of the basal Spotted Ridge formation are found a few feet above this horizon at some localities. The uplift appears to have been epeirogenic, unaccompanied by any notable warping, and agrees with the emergence widely recognized in America near the end of Mississippian time. Whether extensive stripping or channeling preceded the initiation of Spotted Ridge deposition is not known.

In northern California the upper part of the Baird formation is approximately equivalent to the Coffee Creek. According to Diller (1906, p. 3) much tuffaceous material is present in both lower and upper Baird indicating extensive volcanism. No evidence of such activity has been found in central Oregon, though more careful petrographic study might reveal pyroclastic debris.

Localized and very coarse boulder conglomerates within the Spotted Ridge Pennsylvanian bear witness to a high standing and probably rugged terrain not far distant. These boulder beds are intimately associated with plant-bearing deposits but in places extend also into marine facies of the formation. Occurrence of marine or brackish-water shells with land plants emphasizes the probability that we are dealing here with near-shore delta-plain sediments extending seaward from a highland belt drained by streams of great carrying power during flood. Crustal instability with oscillation from subaerial to estuarine or marine conditions is indicated. Swampy areas on the delta flood plain were but temporary sites of vigorous plant growth. In the fine mudstone lenses entire fern fronds are unusually well preserved and show no evidence of prolonged transportation. Calamite stems oblique to bedding are probably in position of growth.

In general the environmental conditions of Spotted Ridge time were analogous to those of the Pottsville in other American coal basins, where following post-Mississippian uplift coal swamps came into being along the flanks of elevated belts (White, 1925). However, in Oregon the land was either not sufficiently stable or sedimentation was too rapid to allow formation of true coal beds. The present area was probably very close to the upwarped axis. Farther away, perhaps to the west.

and more remote from the belt of conglomerate accumulation, conditions may have been more favorable for coal formation.

An unusually high percentage of basic igneous boulders in conglomerates of the Spotted Ridge shows that large areas of these rocks were being eroded. Whether this igneous phase long antedated Spotted Ridge time cannot be determined. A relationship with the Baird eruptive episode in the Lower Carboniferous of northern California does not, however, appear entirely unfounded. In the Baker area to the east there is evidence of Pennsylvanian igneous activity (Gilluly, 1937, p. 15). While there appears to be no direct indication of such activity in this area bedded cherts of the Spotted Ridge may be related to the same general igneous phase.

Marine withdrawal at the close of Baird Lower Carboniferous in northern California conforms to emergent conditions during the Spotted Ridge interval. So far as known there is no depositional equivalent of the Spotted Ridge between the Baird formation and the McCloud Permian in California.

Crustal instability of Pennsylvanian time in central Oregon culminated in mild warping of the Spotted Ridge beds, for angular discordance is recognized between this formation and the Coyote Butte Permian. Limestones near the base of the Coyote Butte formation contain a high percentage of pebbles and arenaceous debris probably derived from the underlying Pennsylvanian. While limestone masses form the best outcrops in the Permian of this region it is probable that sandy and conglomeratic facies constitute a large part of the division. The limestones often appear at the surface as lenslike bodies which may pass marginally into a limy sediment containing much arenaceous debris. It is suggested that some of these prominent masses represent local marine situations of the bioherm type where growth of corals, fusulinids, and crinoids was for a time particularly favored. Asiatic affinity of the fauna indicates a continuous marine connection with that part of the world during Coyote Butte time. While there is evidence of Permian volcanic activity in the Baker area, in northern California and in Nevada (Wheeler, 1939), such has not been recognized in the central Oregon Coyote Butte interval.

Paleozoic history in this portion of central Oregon ends with pronounced folding of the Coyote Butte formation and older rocks. Triassic conglomerates and sands rest upon the Permian beds with marked angular unconformity particularly well demonstrated by discordance in strike. Recognition of this deformation is significant in view of general scarcity of established evidence for folding at the end of Paleozoic time in western North America. Subsequent compression has no doubt greatly accentuated steepness of the earlier folds. Triassic and Jurassic rocks of the

same region stand at high angles, though in the area of this survey they are on the whole less sharply inclined than the Paleozoics. According to Lupher (1941, p. 228) the steeply folded Jurassic sediments in the surrounding country show a wide divergence of fold axes from the trends of Triassic flexures, indicating at least a possible three-fold compression of this highly disturbed region since Coyote Butte Permian time. Evidences of intense dynamic metamorphism of Paleozoic rocks so well shown in the Baker area are not found in this part of central Oregon.

REFERENCES CITED

Buddington, A. F., and Chapin, Theodore (1929) *Geology and mineral deposits of southeastern Alaska,* U. S. Geol. Survey, Bull. 800.

Crockford, M. B. B., and Warren, P. S. (1935) *The Cache Creek series of British Columbia,* Royal Soc. Canada, Tr. (3), vol. 29, sec. 4, p. 149-161.

Diller, J. S. (1906) *Description of the Redding quadrangle,* U. S. Geol. Survey, Atlas folio no. 138.

——— (1909) *Mineral resources of the Grants Pass quadrangle and bordering districts, Oregon,* U. S. Geol. Survey, Bull. 380, p. 48-79.

——— (1914) *Mineral resources of southwestern Oregon,* U. S. Geol. Survey, Bull. 546.

Gilluly, James (1937) *Geology and mineral resources of the Baker quadrangle, Oregon,* U. S. Geol. Survey, Bull. 879.

Gonzales, Norman B. (1933) *A contribution to the paleontology of the Paleozoic faunas of Central Oregon,* Ore. St. Coll., Master's thesis.

Hinds, N. E. A. (1932) *Paleozoic eruptive rocks of the southern Klamath Mountains,* Univ. Calif. Publ., Bull. Dept. Geol. Sci., vol. 20, no. 11, p. 375-410.

——— (1933) *Geologic formations of the Redding-Weaverville districts, northern California,* Calif. St. Div. Mines, Rept. 29 St. Mineral., p. 76-122.

Lupher, R. L. (1941) *Jurassic stratigraphy of central Oregon,* Geol. Soc. Am., Bull., vol. 52, p. 219-270.

McKitrick, W. E. (1934) *The geology of the Suplee Paleozoic series of Central Oregon,* Ore. St. Coll., Master's thesis.

Merriam, C. W. (1942) *Carboniferous and Permian corals from central Oregon,* Jour. Paleont., vol. 16, no. 3, p. 372-381.

———, and Berthiaume, S. A. (1940) *Late Paleozoic history of central Oregon,* (Abstract) Geol. Soc. Am., Bull., vol. 51, p. 1935.

Moffit, F. H. (1938) *Geology of the Chitina Valley and adjacent area, Alaska,* U. S. Geol. Survey, Bull. 894.

——— (1938) *Geology of the Slana-Tok District, Alaska,* U. S. Geol. Survey, Bull. 904.

Moore, B. N. (1937) *Nonmetallic mineral resources of eastern Oregon,* U. S. Geol. Survey, Bull. 875.

Packard, E. L. (1928) *A new section of Paleozoic and Mesozoic rocks in central Oregon,* Am. Jour. Sci., 5th ser., vol. 15, p. 221, 224.

——— (1929) *Discovery of the Baird Mississippian fauna of central Oregon,* (Abstract) Geol. Soc. Am., Bull., vol. 40, p. 257.

——— (1932) *A contribution to the Paleozoic geology of central Oregon,* Carnegie Inst. Washington, Pub. 418, p. 105-113.

Read, C. B., and Merriam, C. W. (1940) *A Pennsylvanian flora from central Oregon,* Am. Jour. Sci., vol. 238, p. 107-111.

Smith, J. P. (1894a) *The Arkansas coal measures in their relation to the Pacific Carboniferous province,* Jour. Geol., vol. 2, p. 187-204.

——— (1894b) *The Metamorphic series of Shasta County, California,* Jour. Geol., vol. 2, p. 588-612.

Thompson, M. L., and Wheeler, H. E. (1941) *Permian fusulinids from British Columbia, Washington and Oregon,* (Abstract) Geol. Soc. Am., Bull., vol. 52, p. 1984.

Washburne, C. W. (1903) *Notes on the marine sediments of eastern Oregon,* Jour. Geol., vol. 11, p. 224-229.

Wheeler, H. E. (1936) *Stratigraphy and fauna of the McCloud limestone,* (Abstract) Geol. Soc. Am., Pr. 1935, p. 409.

———, (1939) *Permian volcanism in Western North America,* Sixth Pacific Sci. Cong., Pr., p. 369-376.

White, David (1925) *Environmental conditions of deposition of coal,* Min. Metall., no. 1402-I.

Williams, J. S. (1941) *Late Paleozoic faunas of Alaska,* (Abstract) Geol. Soc. Am., Bull., vol. 52, p. 1978.

REGISTER OF LOCALITIES

(Pl. 1)

1.—Triassic conglomerate and sandstone; limestone cobbles with Paleozoic fossils including *Striatifera.*

2.—Coffee Creek formation; type locality with abundant fossils.

3.—Coffee Creek formation; sandy limestone with *Striatifera.*

4.—Conglomerate with large pieces of limestone containing *Striatifera;* probably represents Triassic.

5.—Spotted Ridge formation, marine facies; pebbly sandstone and conglomerate with gastropods, fragments of brachiopod shells, and large crinoid columnal segments; coarse conglomerate a few hundred feet south of this locality.

6.—Spotted Ridge formation; conglomerate with large boulders of plutonics.

7.—Spotted Ridge formation; interbedded sandstone and conglomerate, large boulders with diameter of 2 feet.

10.—Spotted Ridge formation; pebbly sandstone and conglomerate.

11.—Spotted Ridge formation; cross-bedded pebbly sandstone with magnetite sand beds.

12.—Coyote Butte formation; cherty limestone with fusulinids.

13.—Coyote Butte formation; limestone with interbedded black chert.

14.—Coyote Butte formation; cherty limestone.

15.—Coyote Butte formation; cherty limestone with *Parafusulina* and *Schwagerina* cf. *linearis.*

16.—Spotted Ridge formation; pebbly sandstone.

17.—Coyote Butte formation; limestone with *Fusulinella* and *Schwagerina.*

18.—Spotted Ridge formation; pebbly sandstone.

19.—Coffee Creek formation; limestone with *Striatifera.*

20.—Coyote Butte formation; limestone with large brachiopod and coral fauna.

21.—Triassic series; outcrop of cherty rock probably representing silicification from an intrusive stock; limestone cobbles with *Triticites* near by.

22.—Triassic series; pebbly sandstone.

24.—Outcrop of bedded bluish-gray limestone. No fossils found.

25.—Triassic series; pebbly sandstone and conglomerate with limestone boulders containing the Coffee Creek *Gigantella* fauna.

27.—Triassic series; pebbly sandstone and conglomerate.

28.—Bedded chert; probably of Spotted Ridge Pennsylvanian age.

29.—Triassic series; sandstone and conglomerate with limestone cobbles containing *Triticites.*

30.—Coyote Butte formation; cherty limestone body with *Parafusulina;* probably lenticular in sandstones.

36.—Coffee Creek formation; limestone with *Striatifera.*

37.—Coffee Creek formation; limestone with *Striatifera.*

38.—Spotted Ridge formation; fine-grained greenish sandstone with plant remains.

51.—Coyote Butte formation; limestone with *Parafusulina* and *Waagenophyllum.*

52.—Coyote Butte formation; cherty limestone.

53.—Spotted Ridge formation; fine sandstone with plant remains.

54.—Spotted Ridge formation; greenish pebbly sandstone with abundant crinoid columnals.

55.—Coffee Creek formation; limestone with *Striatifera.*

56.—Coyote Butte formation; float boulders with fossils.

57.—Coyote Butte formation; limestone with *Parafusulina.*

58.—Bedded chert; probably of Spotted Ridge age.

59.—Spotted Ridge formation; conglomerate.

60.—Coffee Creek formation; limestone with *Striatifera.*

61.—Coffee Creek formation; limestone with *Striatifera.*

62.—Coffee Creek formation; limestone with *Striatifera* and *Gigantella.*

63.—Coffee Creek formation; limestone with *Striatifera.*

65.—Triassic series; conglomerate with limestone cobbles and boulders, some with *Parafusulina.*

66.—Triassic series; conglomerate with limestone cobbles containing *Striatifera.*

67.—Coyote Butte formation; limestone with *Parafusulina.*

68.—Coyote Butte formation; cherty limestone with *Parafusulina* and *Schwagerina.*

69.—Coyote Butte formation; fossil locality.

70.—Triassic series; limestone boulders with Coyote Butte fauna.

71.—Triassic series; pebbly limestone boulders with *Schwagerina* and *Fusulinella.*

72.—Spotted Ridge formation; heavy conglomerate with plutonic boulders.

73.—Coyote Butte formation; pebbly limestone with *Parafusulina* and *Fusulinella.*

74.—Coyote Butte formation; cherty limestone with *Parafusulina.*

75.—Coffee Creek formation; limestone with fauna.

77.—Coyote Butte formation; pebbly limestone with *Parafusulina* and *Schwagerina.*

78.—Coyote Butte formation; limestone with *Parafusulina.*

79.—Coyote Butte formation; brecciated dolomitic limestone with *Parafusulina.*

80.—Coyote Butte formation; limestone with coral fauna and *Parafusulina.*

82.—Coyote Butte formation; cherty limestone with *Parafusulina.*

83.—Spotted Ridge formation; conglomerate with plutonic boulders.

84.—Spotted Ridge formation; conglomerate with heavy plutonic boulders.

85.—Coyote Butte formation; pebbly limestone.

86.—Spotted Ridge formation; conglomerate with heavy boulders.

87.—Coyote Butte formation; limestone with large fauna including *Parafusulina*.

88.—Coyote Butte formation; cherty dolomitic limestone with *Parafusulina*.

90.—Conglomerate with large plutonic boulders; near contact of Spotted Ridge and Coyote Butte formations.

91.—Spotted Ridge formation; plant remains in fine sandstone.

92.—Spotted Ridge formation at spring; conglomerate with large plutonic boulders.

93.—Coffee Creek formation; limestone with fauna including *Gigantella* and *Striatifera*.

94.—Pieces of limestone with *Parafusulina;* probably from small residual area of Triassic conglomerate.

95.—Spotted Ridge formation; heavy conglomerate with plutonic boulders.

96.—Spotted Ridge formation; plant-bearing sandstone.

97.—Coffee Creek formation; limestone outcrop.

98.—Coffee Creek formation; limestone with *Striatifera*.

99.—Coffee Creek formation; limestone outcrop.

100.—Coffee Creek formation; limestone outcrop.

101.—Coyote Butte formation; coral fauna in limestones adjacent to sandstone and pebble conglomerate.

102.—Coyote Butte formation; limestone with *Parafusulina*.

103.—Coyote Butte formation; northwest side Tuckers Butte near axis of an anticline; bedded limestone with *Parafusulina*.

104.—Coyote Butte formation; limestone with *Parafusulina*.

105.—Unconformable contact of Spotted Ridge sandstones and overlying Coyote Butte limestone on east limb of a syncline.

106.—Coyote Butte formation; cherty limestone with *Parafusulina*.

107.—Spotted Ridge formation; conglomerate.

108.—Spotted Ridge formation; plant locality.

109.—Triassic series; conglomerate boulders with *Striatifera*.

110.—Coyote Butte formation; limestone with fusulinids.

111.—Coyote Butte formation; cherty limestone with *Parafusulina*.

112.—Coyote Butte formation; float boulder with *Striatifera*.

113.—Coyote Butte formation; fossil locality.

115.—Spotted Ridge formation, type locality; best plant-collecting locality in fine mudstones.

116.—Coyote Butte formation; limestone with fusulinids.

117.—Coyote Butte formation; limestone with fusulinids on east limb of overturned syncline, dip to east.

118.—Coyote Butte formation; limestone with fusulinids and Coyote Butte fauna.

119.—Triassic series ?; fauna in fine calcareous sandstone.

120.—Coyote Butte formation; limestone with fusulinids.

121.—Coyote Butte formation; limestone with fusulinids.

122.—Coffee Creek formation; limestone with *Striatifera*.

123.—Coyote Butte formation; limestone with *Parafusulina*.

124.—Coffee Creek formation; limestone with *Striatifera*.

125.—Coyote Butte formation; *Schwagerina*.

126.—Coffee Creek formation; limestone with *Striatifera*.

127.—Coyote Butte formation; dolomitized and brecciated limestone.

128.—Coyote Butte formation; cherty limestone with *Parafusulina*.

130.—Coyote Butte formation; cherty fusulinid limestone badly brecciated; *Parafusulina.*

131.—Coffee Creek formation; unfossiliferous limestone.

132.—Coffee Creek formation; limestone with *Striatifera.*

133.—Coyote Butte formation; cherty limestone with *Parafusulina.*

134.—Coyote Butte formation; cherty limestone with *Parafusulina* and *Schwagerina.*

135.—Coyote Butte formation; limestone with *Parafusulina* and corals.

137.—Coyote Butte formation; fusulinid limestone in section 33; *Parafusulina* and *Fusulinella.*

138.—Coyote Butte formation; limestone with *Parafusulina.*

139.—Coyote Butte formation; limestone with *Parafusulina.*

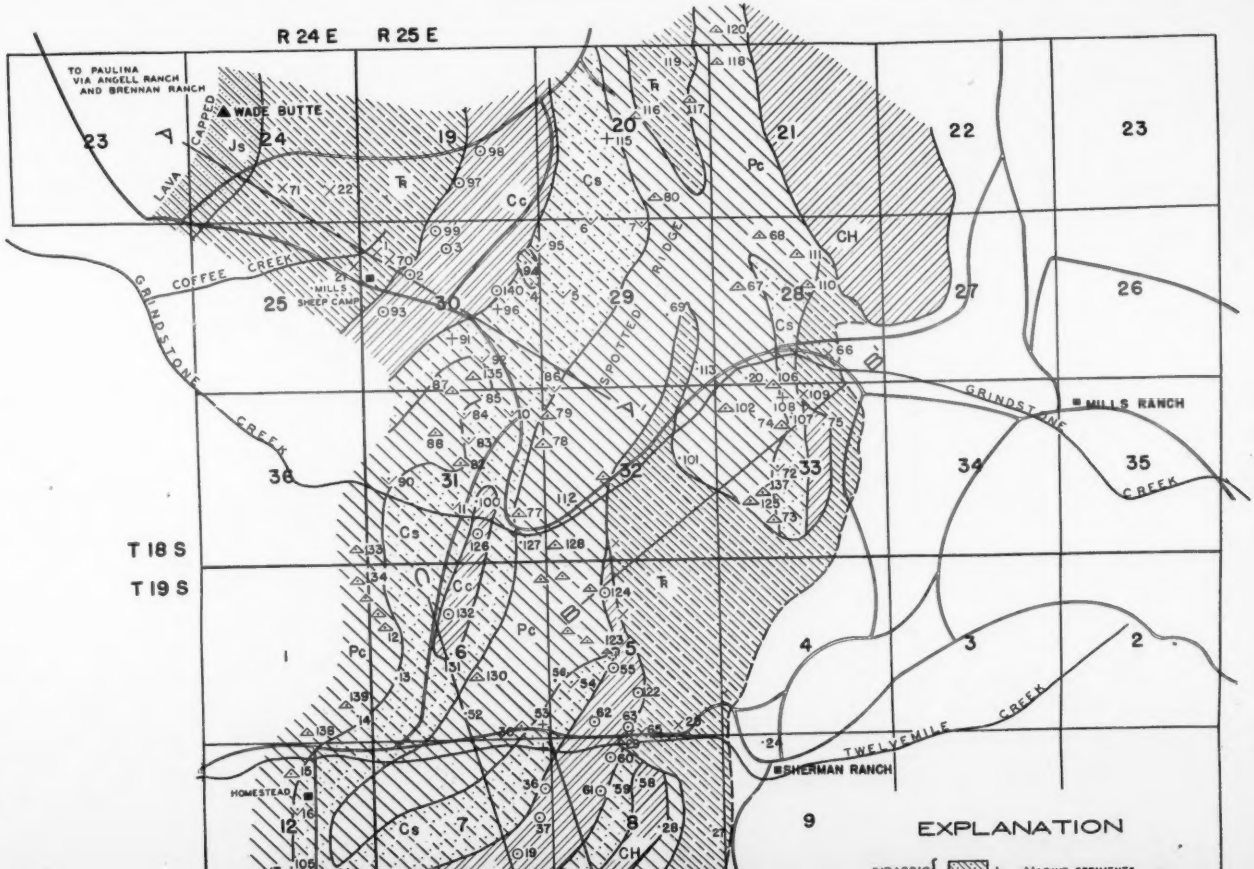
140.—Coffee Creek formation; limestone with *Striatifera.*

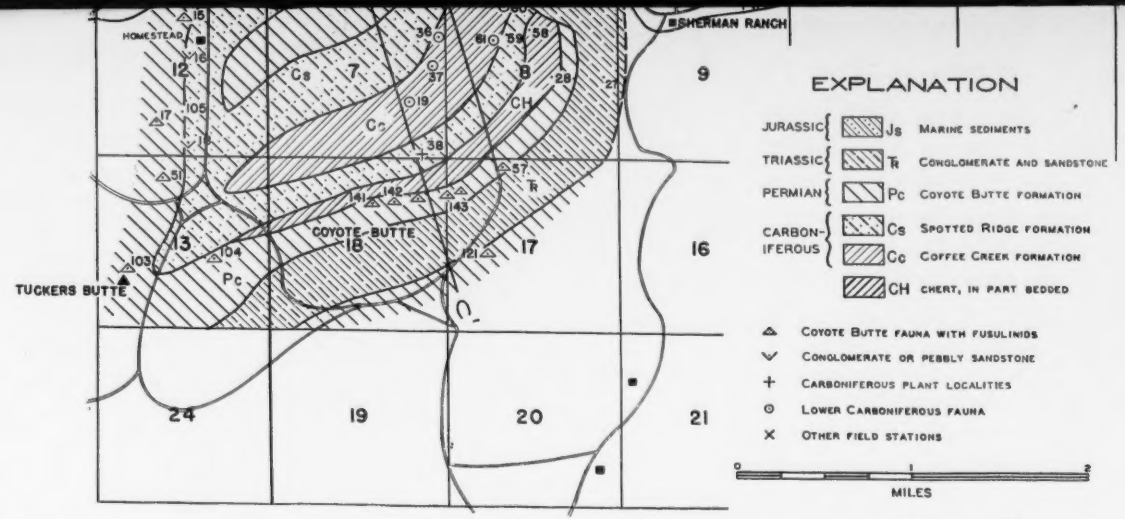
141.—Coyote Butte formation; fusulinid limestone; *Parafusulina.*

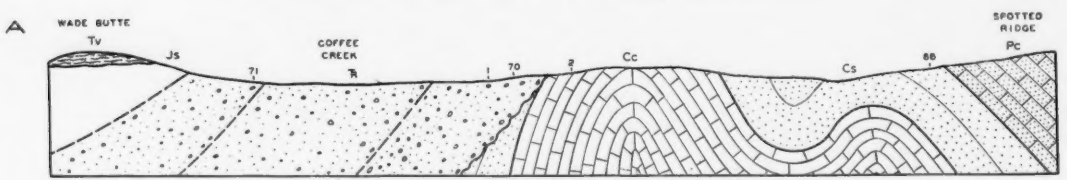
142.—Coyote Butte formation; fusulinid limestone; *Parafusulina.*

143.—Coyote Butte formation; fusulinid limestone near east end Coyote Butte; *Parafusulina, Schwagerina* and *Fusulinella.*

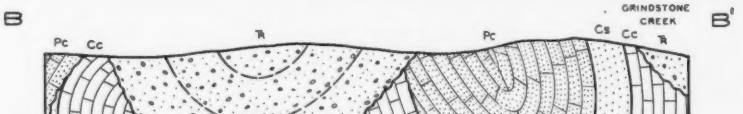
CORNELL UNIVERSITY, ITHACA, N. Y.; THE TEXAS COMPANY, HOUSTON, TEXAS.
MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, MARCH 13, 1942.
PRESENTED BY TITLE BEFORE THE SOCIETY, DECEMBER 28, 1940.


R 24 E
R 25 E
T 18 S
T 19 S
TO PAULINA VIA ANGELL RANCH AND BRENNAN RANCH
WADE BUTTE
CAPPED LAVA
COFFEE CREEK
GRINDSTONE CREEK
MILLS SHEEP CAMP
SPORTED RIDGE
MILLS RANCH
TWELVEMILE CREEK
SHERMAN RANCH
HOMESTEAD
EXPLANATION



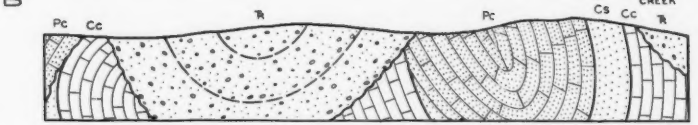


SECTION ALONG LINE A-A'



EXPLANATION OF LETTER SYMBOLS

- Tv TERTIARY VOLCANICS
- Js JURASSIC MARINE SEDIMENTS
- Tr TRIASSIC CONGLOMERATE AND SANDSTONE
- Pc PERMIAN, COYOTE BUTTE FM.
- Cs CARBONIFEROUS, SPOTTED RIDGE FM.

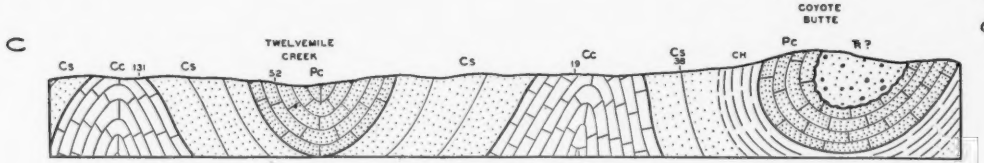


SECTION ALONG LINE B-B'

Js JURASSIC MARINE SEDIMENTS
Tr TRIASSIC CONGLOMERATE AND SANDSTONE
Pc PERMIAN, COYOTE BUTTE FM.
Cs CARBONIFEROUS, SPOTTED RIDGE FM.
Cc LOWER CARBONIFEROUS, COFFEE CREEK FM.
CH CHERT

0 1000 2000 4000
FEET

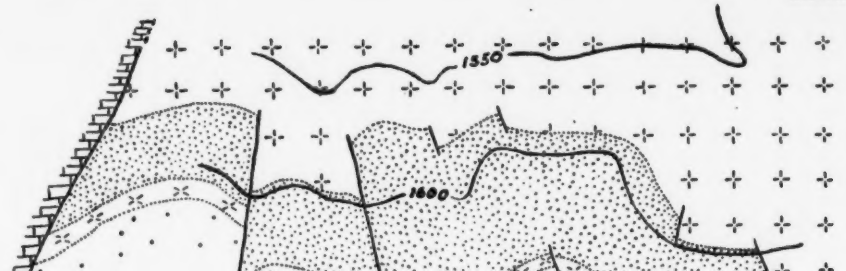
VERTICAL SCALE NORMAL

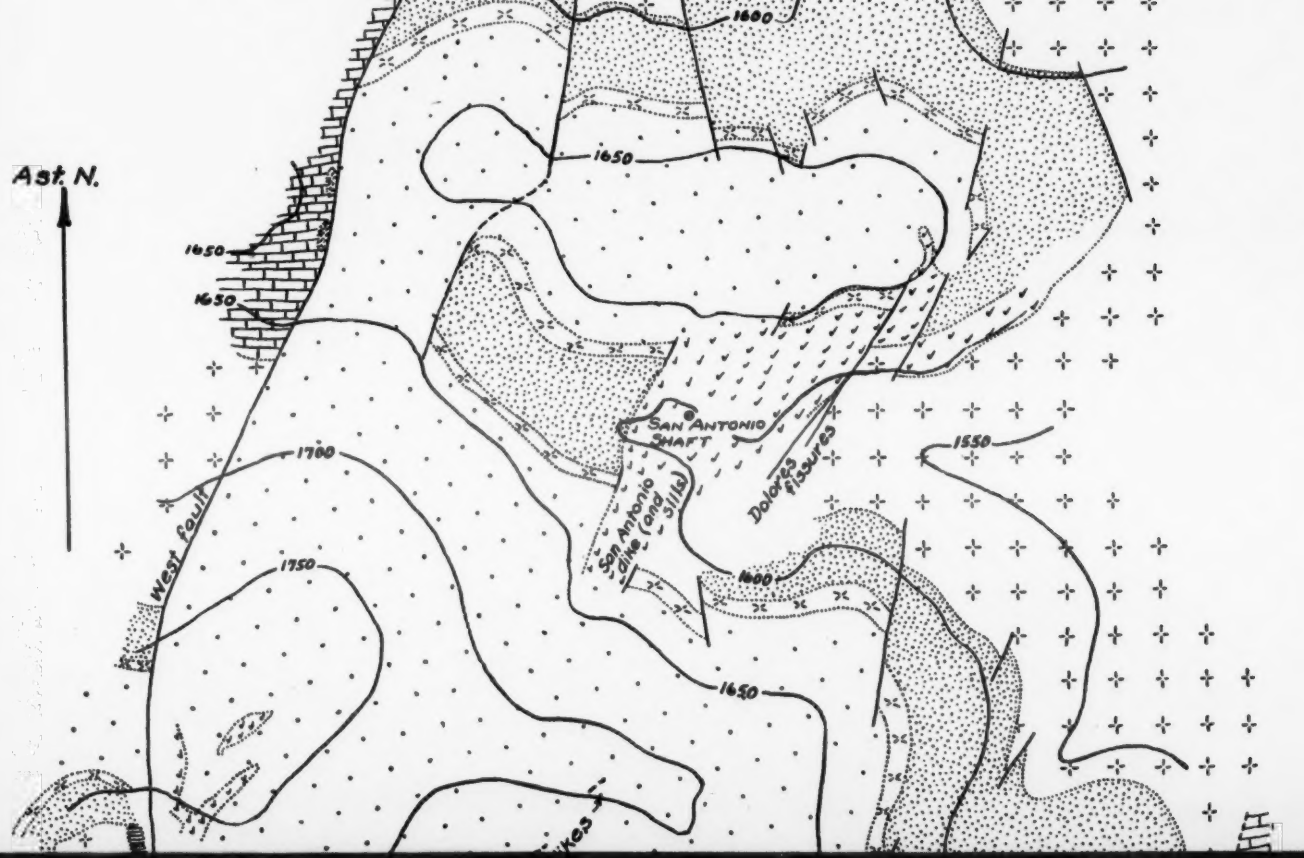


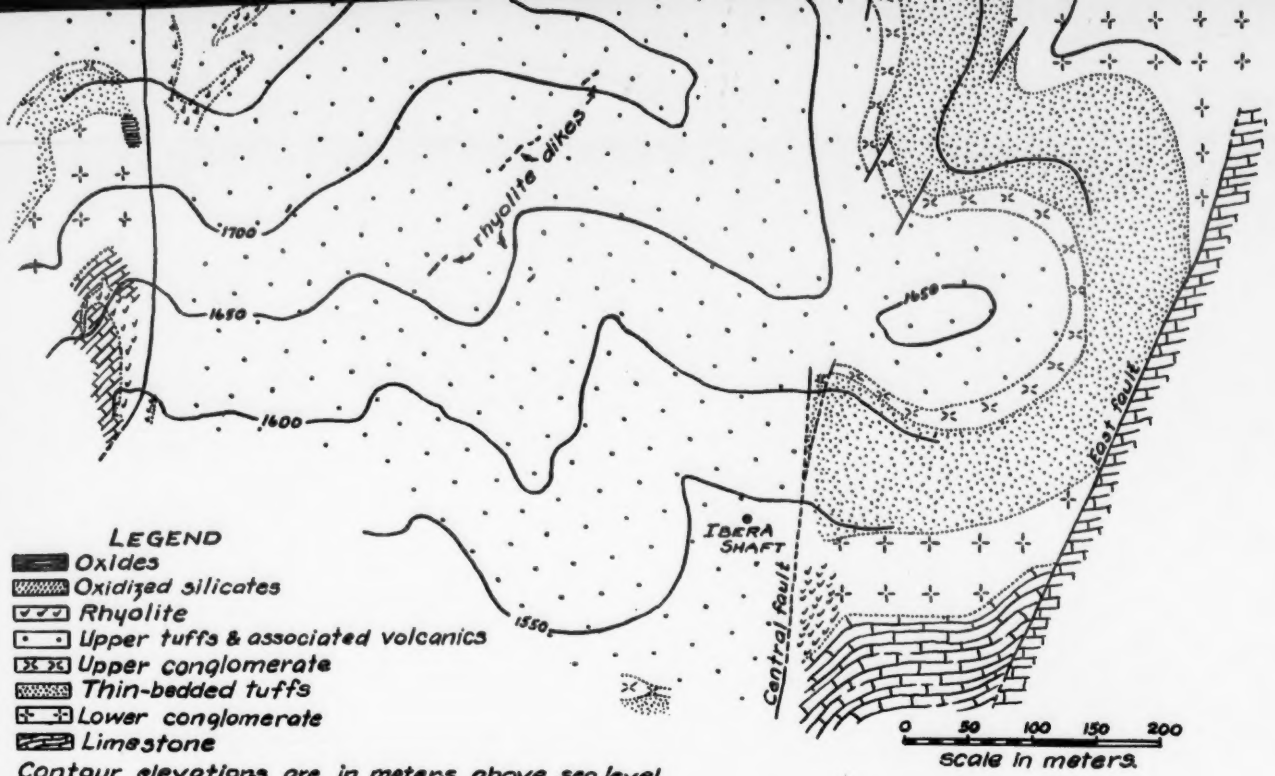
SECTION ALONG LINE C-C'

GEOLOGIC RECONNAISSANCE MAP AND SECTIONS OF GRINDSTONE-TWELVEMILE CREEKS PALEOZOIC AREA, CENTRAL OREGON.

Several small areas of volcanic cap rock omitted.


1550
1600


Ast. N.
1650
1650
1650
1700
1750
1600
1650
1550
West fault
San Antonio shaft
San Antonio dike (and sills)
Dolores fissures



GEOLOGIC SURFACE MAP OF PART OF THE SAN ANTONIO GRABEN

Santa Eulalia district, Chihuahua

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 54, PP. 173-204, 2 PLS., 7 FIGS. FEBRUARY 1, 1943

GEOLOGY AND MINERALIZATION OF THE SAN ANTONIO MINE, SANTA EULALIA DISTRICT, CHIHUAHUA, MEXICO

BY WILLIAM PAXTON HEWITT

CONTENTS

	Page
Abstract	174
Introduction and acknowledgments	174
General setting of Santa Eulalia district	175
Rock types	177
General statement	177
Sedimentary rocks	177
Limestone series	177
Capping series	178
Intrusive rocks	179
Rhyolite dikes and sills	179
Andesite sills	179
Diorites, monzonites, quartz monzonites	179
The graben	179
San Antonio dike	181
Ore bodies	181
Types	181
Localization	183
Mineralization	184
Contact silicates	184
Contact silicates and sulphide relations	185
Sulphides	185
Accessory minerals	185
Micropegmatites	186
Petrography	186
Tin	187
Field observations	187
Petrographic observations	188
Tin ore bodies	189
General considerations	189
Tin chimney	189
Dolores fissures	193
Cocks ore body	195
Vanadium ores	199
Oxidized zinc ores	200
Origin	201
Summary of conclusions	202
References cited	202

ILLUSTRATIONS

Figure		Page
1.	Outline maps: Republic of Mexico; State of Chihuahua	176
2.	Outline map of Santa Eulalia district	180
3.	Plan of part of the San Antonio 6th level	182
4.	Tin chimney	190
5.	Dolores fissures	192
6.	Longitudinal projection striking parallel to trend of dikes	195
7.	Cocks ore body	196

Plate		Facing page
1.	Geologic surface map of part of the San Antonio graben	173
2.	Photographs of San Antonio ores	204

ABSTRACT

The mineral deposits of the San Antonio mine, in the Santa Eulalia district of Chihuahua, Mexico, occur in Upper Cretaceous and possibly Lower Cretaceous limestones. They lie within a downfaulted limestone block, are closely associated with a zone of acid dikes, and form replacement deposits of the contact-metamorphic type.

The chief yield has been lead and silver ore, but tin and vanadium ores have been produced commercially. The bodies are unusual in that lead and vanadium are important products from a contact-metamorphic area and in that tin occurs in replacement deposits in limestone.

Detailed descriptions of tin and vanadium ore bodies are given. The writer concludes that tin was the last ore element to be introduced and suggests it may have been pneumatolytic; that the vanadium ore has been derived in large measure from the breakdown of primary vanadium-bearing silicates; and that the deposits are best classed in the pyrometasomatic group, of contact origin.

INTRODUCTION AND ACKNOWLEDGMENTS

The San Antonio mine has produced silver, lead, tin, vanadium, and zinc, with copper and gold as by-products. Its deposits, dominantly of the chimney form, resemble the contact-metamorphic type although the metamorphic source has not been found. The mine is unique in being the first highly successful tin producer in North America, in that lead has been the most important product from a deposit of this kind, and in that primary vanadium-bearing silicates may have been the major vanadium ore source.

Data for this paper have been compiled from routine field studies undertaken while in the employ of the American Smelting & Refining Company, and permission for publication has been obtained through the courtesy of the Company. In addition the Company has made available certain petrographic examinations by Dr. Roy D. McLellan and mill reports by Mr. R. J. Mellen.

Also it should be remembered that the geologic knowledge pertaining to any mining area is rarely the result of one man's work, especially in a district as old as this, and that each new student benefits from the work

of his predecessors. It has been the privilege of the author to work in many new and vital sections of the mine, but he acknowledges his indebtedness to all students of the district, operators as well as geologists, and especially to Mr. T. P. Clendenin, Mining Geologist for the Mexican Division of the American Smelting & Refining Company, for his observations on the geology of the area and for his criticisms of the paper. In addition he wishes to thank Professors C. P. Berkey and P. F. Kerr of Columbia University for their guidance and encouragement, and Professor Kerr for his criticisms of the paper.

GENERAL SETTING OF SANTA EULALIA DISTRICT

The mine, in central Chihuahua, 25 kilometers S. 85° E. from Chihuahua City and 7 kilometers N. 70° E. from the village of Santa Eulalia (Fig. 1), is part of the Santa Eulalia mining district and lies in a range near the western edge of the Mexican plateau's limestone province. This range, rising 650-700 meters above adjacent valleys, is composed of limestone thinly mantled with a limestone conglomerate of variable thickness. Overlying the conglomerate are andesitic and rhyolitic volcanics. The volcanics have no economic interest, and the conglomerate rarely has. The limestone, however, is widely and strongly mineralized.

The structure of the range is difficult to determine because of the absence of key horizons. The beds are horizontal or slightly tilted, and general field observations indicate a dome about 8 kilometers in diameter whose eastern and western flanks are truncated, apparently by faulting.

The topography is rugged, and the stream courses form both open canyons and deep, narrow, straight-walled gorges approaching 250-meter depths. In some of these are remnants of small terraces. An older and extremely rugged surface underlies the volcanics and conglomerates, and its buried valleys are probably as deep and steep-walled as those being formed at present.

The district is divided into eastern and western areas, geologically distinct and separated by about 5 kilometers of barren limestone (Fig. 2). Limestones in the western area dip southwest; faults are small and insignificant; intrusives are almost entirely sills; ore bodies are dominantly of the manto type; ores are simple mixtures of galena, marmatite, pyrite, and pyrrhotite; tin and vanadium are not found; and lime-iron silicates are rare, but where they occur they form large homogeneous masses of ilvaite and fayalite. This area, described by Prescott (1916), has been exploited since the days of the Spaniards and has been enormously important.

FIGURE 1.—*Outline maps: Republic of Mexico and State of Chihuahua*

Limestones in the eastern area dip southeast, and have been tilted by numerous displacements; major faults are important; intrusives are mainly dikes; ore bodies are almost entirely chimneys; ores, while occasionally simple sulphides, are in many places an intricate mixture of sulphides and such lime-iron silicates as garnet, tremolite, and actinolite; and tin and vanadium are important products. In this area San Antonio is the only important mine.

The mine lies in a steep-walled northerly trending valley floored by the Capping series and walled by limestone (Pl. 1). About 3 kilometers long by 0.5—1.0 kilometer wide, the valley is obstructed to the northwest of its center by a peninsulalike spur which swings in from the west wall. Within this spur is a northerly trending rhyolite dike, and nearby is the San Antonio mine. The mine workings are confined within the valley walls, generally close to the rhyolite. There are two reasons for this: (1) The valley reflects a graben, and (2) the dike, or dikes, have localized the ore bodies.

ROCK TYPES

GENERAL STATEMENT

The formations in the San Antonio area include sedimentary rocks and intrusives. Volcanic rocks are intimately associated with the sediments, and for convenience are grouped with them.

Sedimentary rocks:
- Limestone series
- Capping series

Intrusive rocks:
- Rhyolite dikes and sills
- Andesite sills
- Diorites, monzonites, and quartz monzonites.

SEDIMENTARY ROCKS

Limestone series.—This coarsely bedded, light-gray, dense, and monotonously uniform series has been prospected in San Antonio through a vertical range exceeding 300 meters, and in the western district for over 900 meters without revealing the base. It carries few cherts and few fossils, and no guide fossils have been recognized. Its exact age is not known, although a suite of fossils collected 200 meters below the precapping erosion surface indicates Upper Cretaceous age. These fossils, identified by Dr. H. N. Coryell of Columbia University (personal communication) and to be described by him elsewhere, consist of new specific forms of *Peregrinella* sp., *Kingenia* sp., and *Cyclothyris* sp., all of Upper Cretaceous age; together with segments of Octocoralla and crinoid stems, and fragments of high-spired gastropods, from which no genera or species were identifiable.

However, these limestones are the same as those in the western district which Spurr, in a Company report of 1911, refers to as a "Mesozoic (probably Cretaceous) limestone." Prescott (1916) describes them as "a Cretaceous limestone of unknown thickness" and states "the oldest . . . formation exposed is probably Lower Cretaceous." T. W. Stanton (personal communication to W. P. Popenoe, 1931) tentatively identified a suite of fossils collected 400 meters below the pre-capping erosion surface as of Glen Rose age. He was unable to make specific identifications, and his generic assignments of the brachiopods were tentative. Besides corals and sponges of undetermined genera, the fossils included:

Echinoid spines
Cidaris ? sp.
Goniopygus sp.
Ostrea (*Abetryonia*) sp. cf. *Ostrea carinata* Lamarck
Kingenia sp.
Acanthothyris ? sp.

Since some of these occur at Los Lamentos, 250 kilometers north of Santa Eulalia, in beds containing *Orbitolina texana*, a Glen Rose species, Stanton thought the Santa Eulalia suite might come from Glen Rose beds. There is a marked lithologic similarity between the Santa Eulalia beds and the Aurora limestone (Kelly, 1936, p. 1027), part of which is a Glen Rose equivalent (Kellum, Imlay, and Kane, 1936, p. 1002).

Capping series.—This series of variable thicknesses of conglomerates, tuffs, and flows unconformably overlies the Limestone series. It is widespread, but its subdivisions are of variable composition and local significance. In the San Antonio area the four main members are: Upper tuffs and associated volcanics, Upper conglomerate, Thin-bedded tuffs, and Lower conglomerate. Only the Lower and Upper conglomerates are of interest, the Lower conglomerate being the host rock for small but important ore bodies, while the Upper conglomerate is the one reliable marker horizon.

The Lower conglomerate, not everywhere present, has a maximum thickness of over 100 meters. It is composed of limestone pebbles cemented by a calcareous matrix which in places is tuffaceous. Small pebbles of igneous rock are found, but never in abundance. Scattered irregularly through the formation are sill-like bodies of a light-gray, fine-grained, usually calcareous rock thought to be water-sorted tuffs.

Conformably overlying the Lower conglomerate and grading into it are the Thin-bedded tuffs, approximately 30 meters thick. They consist of thinly bedded rhyolitic and andesitic tuffs and occasional interbedded rhyolite sills or tuffaceous flows.

The Upper conglomerate conformably overlies the Thin-bedded tuffs. It is from 2-6 meters thick and is composed of limestone pebbles cemented by a calcareous tuffaceous matrix.

The Upper tuffs and associated volcanics are the most widespread of the series, and, while they appear to lie conformably above the Upper conglomerate, they rest in some places directly on limestone. Their maximum thickness exceeds 250 meters. They are heavy-bedded, massive, predominately rhyolitic, and are cut by and interbedded with similar flows and sills and dikes.

INTRUSIVE ROCKS

Rhyolite dikes and sills.—Rhyolites are white, dense and variable in hardness and show flow lines and spherulites only in outcrops. They occur as a series of dikes, collectively called the San Antonio dike, and occasionally as sills. One dike cuts the Upper tuffs; a second terminates at the top surface of the Upper conglomerate, as though eroded prior to the deposition of the Upper tuffs. They are the youngest and most important intrusives.

Andesite sills.—In the upper levels, in both limestone and Lower conglomerate, are fine-grained, severely altered andesite sills. Individual exposures have not exceeded 15 meters in thickness. Possibly related are fine- to medium-grained mediumly basic sills outcropping in limestone near the Lower conglomerate. One outcrop is cut by the East fault.

Diorites, monzonites, quartz monzonites.—At the south end of the mine, entirely in limestone, is a greenish, medium- to coarse-textured prerhyolite sill 70 meters thick. It is classified as a diorite and is known as the Ibera sill. In the same vicinity Dr. R. J. McLellan identified altered and oxidized rocks as monzonites, quartz monzonites, and augite diorites. Whether these are altered facies of the Ibera sill or separate intrusives is not known. However, several kilometers to the northwest there is a porphyritic quartz monzonite dike in the Upper tuffs (Fig. 2).

THE GRABEN

The graben strikes approximately north-south, is roughly 500-1000 meters wide, and has been dropped between 150 and 250 meters (Fig. 2; Pl. 1). Its western wall consists of a single normal fault of steep eastward dip, called the West fault. The eastern wall is more complex. It is formed by two normal fault scarps of steep westward dip, the East and Central faults, which converge in hingelike manner near the center of the wall, but never join. The East fault forms the northern part, and its throw increases rapidly in that direction, while the Central fault forms

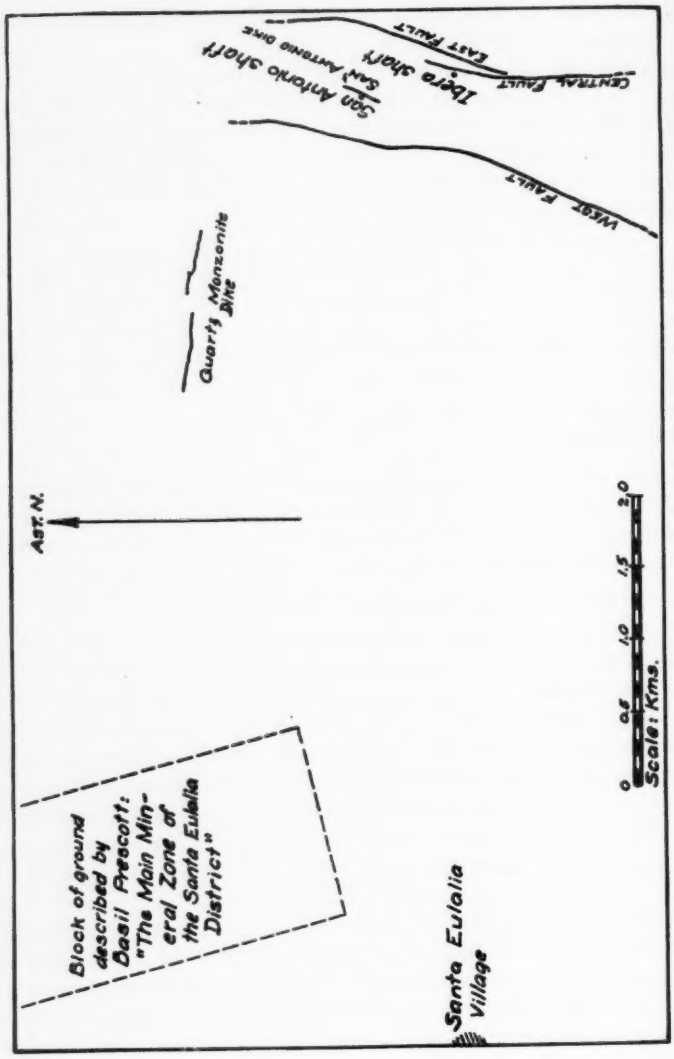


FIGURE 2.—*Outline map of Santa Eulalia district*

the southern extension, and its throw increases rapidly to the south. Within the hinge area the two faults overlap, the East fault veering eastward from the Central fault while the Central fault continues northward on its original course and eventually disappears.

In addition to these faults, whose scarps form the valley walls, there are lesser ones throughout the central portion of the valley that are recognized by displacements in the Upper Conglomerate. Some acted as channels for ore solutions and formed replacement veins, while one of steep dip has been followed by a rhyolite dike.

These are the only structural features recognized. Stratigraphic evidence is so fragmentary that although the beds seem to have a general south-southeast dip no definite attitude has been established.

SAN ANTONIO DIKE

Somewhat to the north and west of the valley's geographical center a rhyolite is exposed. At the surface it is extensively developed as a sill, although actually it is the San Antonio dike which normally is 4-10 meters wide (Pl. 1). Underground work showed it to persist with general northeast strike for 1½ kilometers. To the southeast and en echelon with it is a second dike whose northeast termination lies a few score meters north of the south end of the San Antonio dike. The "noses" or terminations of both dikes pitch northward. Seen in vertical section they overlap by 10 to 30 meters (Figs. 6, 7). These relationships, shown at the 250-meter level by Figure 3, persist downward to a depth of at least 400 meters and persist upward, as shown by widespread prospecting, into the upper levels.

Additional development has disclosed as many as four parallel dikes within a zone less than 100 meters wide. This is attributed to a tendency of the rhyolites to branch as they travel upward. These branches may continue as major dikes or may terminate abruptly. Thus, what appears to be a single dike in surface exposures may be part of a dike zone.

Although the original San Antonio dike does not deviate in strike, the trend of the dike zone and the other individual dikes changes frequently (Fig. 3). The dikes within the zone usually parallel one another except where the zone is changing strike, and in such areas the dikes follow either the old or the new strike.

ORE BODIES

TYPES

In Mexican limestone districts "manto" and "chimney" refer to pipe-like replacement ore bodies of roughly circular cross section in which the ratio of the maximum width to the length of the body is small. A manto

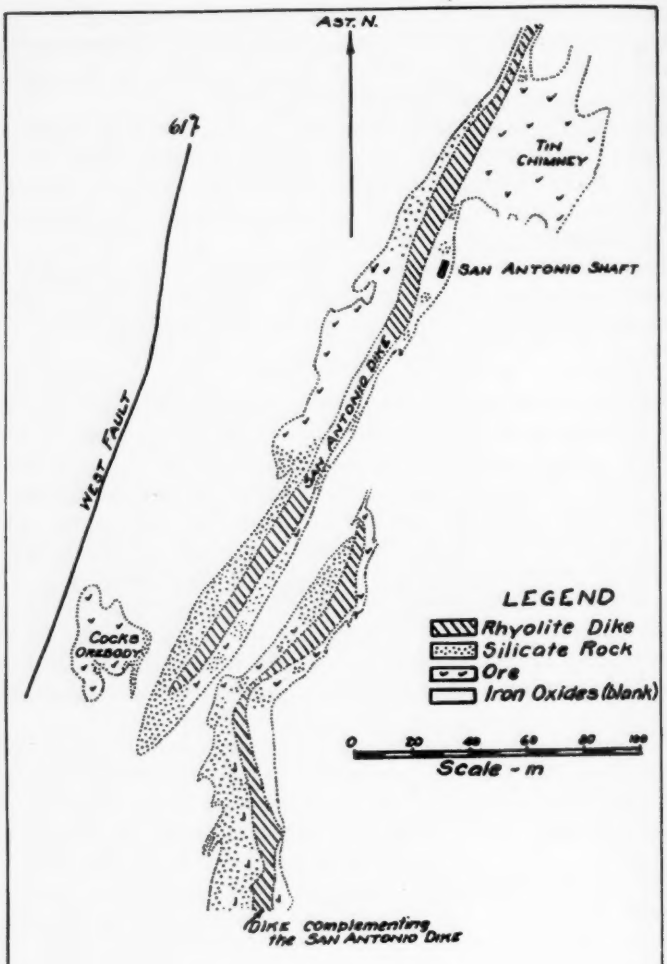


FIGURE 3.—*Plan of part of the San Antonio 6th level*

is a horizontal body of this type, while a chimney is vertical (Prescott, 1926).

There are inclined chimneys and inclined mantos, and in an effort to distinguish between them some authors confine the word manto to pipelike bodies tending to parallel the bedding, and chimney to those bodies which tend to cut the bedding. Thus, they speak of flat-lying chimneys and vertical mantos. Such usage, however, is contrary to general acceptance, and in this paper the word manto is applied solely to bodies more or less horizontal, and chimney to bodies more or less vertical.

In San Antonio most ore has been mined from chimneys, although replacement veins, known as fissures, have furnished important tonnage. Mantos have been small and scarce and usually have sprung from the larger chimneys. There are also widespread masses of mineralization found along the contacts of dikes and sills.

LOCALIZATION

The rhyolite dikes form the principal locus of the most intense and extensive mineralization yet found. However, the contact relationship is accidental, as it can be demonstrated that the rhyolites were not the source of the mineralization, because:

(1) Some rhyolite dikes have clean limestone contacts, and megascopically neither the limestone nor the rhyolite shows signs of alteration.

(2) Rhyolite dikes are sometimes cut by sulphide veinlets whose walls are altered and impregnated with disseminated sulphides, yet away from these veinlets the rhyolite is fresh and lacks disseminated sulphides.

(3) Some rhyolite borders are so severely metamorphosed that it is impossible to locate the original contact or to determine what was originally limestone and what rhyolite, yet the interior of the dike is relatively fresh and unaltered.

(4) Some exposures of lime-iron silicates are isolated from known intrusives.

Possibly the intrusion of the dikes further prepared the ground for the mineralizers. Their intrusion may have released mineralizers within a magmatic reservoir, and these mineralizers, producing the contact metamorphism, merely used pre-existing dike contacts as convenient highways for transportation. Finally, their branching habit may have exerted a control upon the location of individual ore bodies, as is illustrated in the discussion of the Tin chimney.

Other, subordinate controls of the paths of the mineralizers are the diorite and andesite sills. The contacts of these intrusives have been mineralized, but the mineralization seldom has formed ore bodies except at intersections with mineralized fissures.

Where mineralized fissures intersect sills ore bodies frequently formed along the intersection, even though both sill and fissure may lack ore elsewhere. Some sills allowed the mineralizers to continue through, for the formation of additional ore bodies at higher horizons, either in the fissure proper or at an intersection with another sill.

MINERALIZATION

Contact silicates.—San Antonio ore bodies, with the exception of the replacement veins, consist of varying proportions of skarn minerals and ore. Sometimes the commercial minerals are disseminated throughout a silicate gangue; sometimes they occur as a core of iron and lead oxides, with minor silicates, surrounded by a silicate shell; and sometimes the silicate minerals are only small included patches.

Common gangue minerals are garnet, epidote, and fibrous radiating silicates. The garnets, ranging from green to a dirty yellow, are thought to be andradite and grossularite. They occur in fine-grained masses, and crystal faces are small and uncommon. Epidote is massive, somewhat granular, and seemingly fine-grained. Individual crystals have not been recognized, but cleavage faces are occasionally noted. The fibrous silicates—tremolite, actinolite, and possibly hedenbergite—range from minute spherules to coarse nodules 2½ inches in diameter. While intimately associated with other lime-iron silicates, they occasionally form large essentially pure masses of interlacing fibers. The uncommon mineral ilvaite has been observed.

These minerals are found on rhyolite-limestone contacts and in limestone isolated from known rhyolites. In the lower levels epidote is found at the contact with the dikes, with garnet adjoining unaltered limestone. At higher elevations hedenbergite intervenes between the limestone and garnet, the sequence then being rhyolite dike, epidote, garnet, hedenbergite, limestone. Actinolite and tremolite are found in the middle, frequently between the garnet and epidote, while the hedenbergite zone is sometimes an epidote, garnet, hedenbergite mixture. Lindgren (1928, p. 796) describes a similar dike, epidote, garnet, unaltered limestone occurrence from the Morenci district in Arizona.

In bodies isolated from rhyolites there is less zoning and more homogeneity. Hedenbergite tends to lie next to limestone, while epidote concentrations have not been noted. The bodies occur as small chimneys or occasionally as small veins, some of which have connections to the contact.

Contacts between silicates and limestone are sharp and involve a megascopically unaltered limestone. Contacts between silicates and dikes, however, are sometimes severely altered. Where they are changed to epidote the position of the original contact is generally destroyed; where the

contacts are kaolinized the dike is changed to a whitish paste frequently impregnated with sulphides and fluorite.

Contact silicates and sulphide relations.—Hand specimens of unoxidized silicates show sulphides, chiefly marmatite, disseminated as small grains. Others show irregular streaks of marmatite, galena, and pyrite cutting garnet, epidote, and actinolite. These small features are not typical of San Antonio sulphide bodies but are believed to indicate possible time relations between sulphide introduction and silicate mineralizations.

Sulphides.—San Antonio sulphides can be placed in three general groups. The commonest is a coarse dissemination of marmatite in a silicate matrix, while within the marmatite field are occasional isolated veinlets of pyrite and galena. The bodies range from areas dominantly marmatite to areas dominantly silicate, and while generally there is a marked change from sulphide to nonsulphide-bearing matrix the bodies would show "assay walls" if they were of economic interest.

Sometimes these disseminated bodies have merged into massive marmatite, and rarely they are enclosed by limestone. A possible subdivision includes massive, coarsely grained, heterogeneously mixed galena and marmatite bodies which have been found only within limestone walls and show no signs of silicates.

The third group differs from the preceding and is less common. It consists of massive, more or less homogeneous mixtures of galena, marmatite, and pyrite that show no signs of silicates, and which approach the so-called "normal" sulphides of Santa Eulalia's West Camp.

This division of the sulphides into three groups has been made merely for the sake of description and does not imply three distinct phases of sulphide emplacement. There is little or no evidence either favoring or opposing successive stages, and the writer favors a single sulphide introduction. Whatever differences are found might easily arise during one continuous mineralization.

Accessory minerals.—This group of primary minerals that may have been active in the metallization of the mine are of secondary importance in hand specimens. Only those which can be studied megascopically are included: fluorite, quartz, magnetite, and calcite.

Fluorite is found in all levels that are driven in limestone but has not been noted in the conglomerate. It cuts the lime-iron silicates but is of doubtful age with respect to sulphides. It occurs in sulphide bodies but may be contemporaneous.

Quartz is found at all elevations in the mine but has not been identified within massive sulphides. It cuts the silicates and disseminated sulphides

and is associated with sulphide grains of the same kind that it cuts elsewhere. It forms needlelike penetrations in pre-existing silicates and nests of small crystals in open vugs. It silicifies pre-existing rocks, and veinlets of it cutting disseminated sulphides in the less oxidized mineralizations of the upper levels contain acicular, finely bladed magnetite crystals. Since this type of magnetite, without quartz, penetrates disseminated sulphides the quartz-magnetite combination may be later than some sulphides.

Magnetite is widespread but not abundant; it is found in the lower 250 meters of the mine, cuts disseminated sulphides, and is associated with calcite veinlets in the lower levels and with quartz veinlets at intermediate horizons. It has not been found in massive sulphides.

Calcite is as ubiquitous as quartz or fluorite. It replaces patches of hedenbergite, and where replacement has been incomplete there is a nondescript hybrid mixture shot through with well-formed quartz crystals. It also occurs along ore and limestone walls as massive black and gray bodies rich in iron and manganese impurities. It seems to be later than the lime-iron silicates, of the same age as magnetite, probably of the same generation as some quartz but slightly earlier than other quartz, while judging from magnetite relations it may be later than some sulphides. Probably the solutions were saturated with calcium molecules and precipitated calcite wherever they had a chance.

Micropegmatites.—Veinlets of micropegmatite 1-2 inches wide cut silicate fields both near and away from ore bodies but are not reported cutting ore. Sometimes the micropegmatite does not occur in veins but cuts the skarn minerals in a gashy irregular manner, while in other specimens it floods and apparently replaces sulphide grains of zinc, lead, and iron. It tends to grade into quartz veinlets.

This mineralization is found only in the transition zone between oxides and sulphides, although immediately above this horizon are quartz stringers of approximately the same width. While unrecorded from deeper levels it may be unexposed because of insufficient development.

Although no minerals are identified with the micropegmatite, other than occasional sulphide grains, the intergrowth may represent the final stage of definite mineralization and may be the nearest approach yet found to the intrusive that produced the contact metamorphism.

Petrography.—Except for a little work by Dr. Roy D. McLellan on mill concentrates and a few picked specimens, no petrographic study has been made of San Antonio minerals. A few slides, prepared from gangue minerals, added little to what is known from hand specimens and

field evidence, although feldspar (probably orthoclase) was found in slides from the upper levels. In the mine it is identified only near the bottom.

TIN

FIELD OBSERVATIONS

Although San Antonio has been an important tin mine, tin minerals have never been seen in hand specimens, even though its stopes averaged 1.5 per cent tin, while some ran as high as 4 per cent and in rare cases 5 per cent or over. However, from milling operations it is known that the tin mineral is a fine-grained, brownish-yellow, slightly purplish, crystalline cassiterite.

Tin occurs intimately associated with oxide ore bodies but not with any particular mineral nor type of ore. There are ore bodies barren of commercial tin, or in which it is distributed erratically from top to bottom, while in others it is uniformly dispersed. It is found in chimneys, mantos, veins, and along dike contacts, but each type of ore body has produced examples lacking it. There are tin areas intimately associated with garnets and allied silicates, while others are as intimately associated with fluorite; in contrast, some lack all signs of either fluorite or silicates. The mine has few sulphide bodies. Some carry tin, probably in the form of cassiterite, because assays fail to indicate the presence of stannite.

Attempts to find field criteria indicative of tin areas, although unsuccessful, have yielded the following observations:

(1) Areas with high tin assays sometimes show a noticeable concentration of fluorite. However, fluorite cannot be used as an indicative mineral since many fluorite-bearing areas lack tin and many tin-bearing areas lack fluorite.

(2) In the bottom levels certain areas of silicate concentration, normally barren or low in tin, show higher tin assays in samples that include calcite-magnetite veinlets than do neighboring samples which lack them. Tin content for both types is very low and might be on the order of magnitude of:

0.40 per cent Sn in samples with calcite-magnetite veins;

0.05 per cent Sn in samples lacking them.

Therefore, magnetite may be genetically related to tin, especially since both are primary oxide minerals. However, it is not a tin guide because in silicate fields cut by quartz-magnetite veinlets there is no difference in the tin content of samples with or without the quartz and magnetite. In either case the tin content is only about 0.06 per cent. This lack of relationship is also true of samples in

which magnetite crystals unaccompanied by quartz or calcite penetrate pre-existing mineralizations.

(3) In the sulphides at the bottom of the Tin chimney tin is possibly related to an unidentified metallic to submetallic mineral which is not stannite. Tin content is sometimes, but not always, greater where this mineral penetrates sulphides.

(4) The same sulphides contain siliceous veinlets, probably fine quartz, which at a cursory glance resemble dense, light-colored limestone ribs. The tin content ranges from 1.0 to 4.0 per cent in samples including these veinlets, but from 0.0 to 1.0 per cent in samples lacking them.

PETROGRAPHIC OBSERVATIONS

Petrographic study, by Dr. Roy D. McLellan, of mill concentrates and picked specimens showed that although all the cassiterite is fine-grained it varies greatly. It occurs as individual crystals free from accompanying gangue or is intimately associated with hematite, quartz, topaz, and to a minor extent with magnetite, ilmenite, columbite, muscovite, tourmaline, and calcite. In addition pyrite, pyrrhotite, galena, sphalerite, arsenopyrite, and wolframite were found in the same specimens with cassiterite but not intimately associated. While much of the cassiterite was pure, golden, and translucent, large amounts carried hematite and pyrolusite as darkening and clouding impurities. The hematite in particular seemed to bear a definite relationship to cassiterite, and in certain specimens almost all hematite particles were tin bearing.

Quartz and magnetite are the only minerals of this group easily identified in the field, the magnetite of hand specimens occurring in plates frequently ⅜ inch across. Under the microscope, however, both magnetite and wolframite were generaly finer-grained than accompanying particles of other minerals.

The samples studied came from the lower parts of the mine. No petrographic study has been made of ores from the upper levels nor from the mine in general. However, the author examined a thin section from erratic sulphides found in the upper levels, and cassiterite crystals occurred in both sulphides and gangue. Orthoclase feldspar formed part of the gangue. No plagioclase was seen.

Cassiterite is thus evidently later than the sulphides, although there may have been an uninterrupted sequence of mineralization. Field evidence likewise indicates that tin may have been introduced later than both silicates and sulphides, as suggested by the calcite-magnetite veinlets cutting silicates and by the veinlets of light-colored crystalline material cutting sulphides.

The minerals accompanying the cassiterite are distinctive and by themselves suggest a pneumatolytic facies or possibly a pegmatite. Their size precludes a pegmatitic origin, yet the quartz-feldspar veinlets from the bottom levels approach a graphic intergrowth so characteristic of pegmatites, and these veinlets are definitely later than the lime-iron silicates and some sulphides.

Since the associated minerals include a group commonly accepted to be of pneumatolytic origin, and since all evidence points to an origin later than both silicates and sulphides, it is suggested that the erratic distribution of tin was caused by volatile mineralizers which followed closely upon the sulphide introduction, in part using sulphide channels and in part permeating areas of pre-existing mineralization, both silicate and sulphide.

TIN ORE BODIES

GENERAL CONSIDERATIONS

There have been two main sources of tin production—the Tin chimney and a series of replacement veins grouped under the collective heading of the Dolores fissures, which have been consistent tin carriers from top to bottom. In other sources the tin has been localized in certain limited horizons. Such is the Cocks ore body. In its upper part it had a tin content comparable to that of the Tin chimney or the Dolores fissures, but elsewhere it was nearly barren. Besides these three great sources, now exhausted, additional small ore bodies produced ores of good grade but limited tonnage. The three chief sources, however, are typical of San Antonio tin bodies, and they alone will be described.

TIN CHIMNEY

The Tin chimney, over 300 meters high and with an average diameter of approximately 40 meters, lies against the east side of the San Antonio dike (Fig. 4). Near the chimney bottom this dike pinches out, and immediately under the ore body two new dikes appear. Associated with them is a large body of lime-iron silicates.

Where the projected position of the San Antonio dike has been crosscut on the lowest level only lime-iron silicates were found. These silicates, although dominantly free of sulphides, carry disseminated sulphide grains low in lead and iron, and sometimes areas of marmatite concentration. This mineralization extends upward to the bottom of the dike and thence spreads along its sides, and as it rises it carries fewer silicates and more lead until it is finally ore.

Of the three dikes the most eastern is the one whose top plunges northward. It lies directly beneath the center of the chimney, yet where explored on the bottom level its associated mineralization is extremely

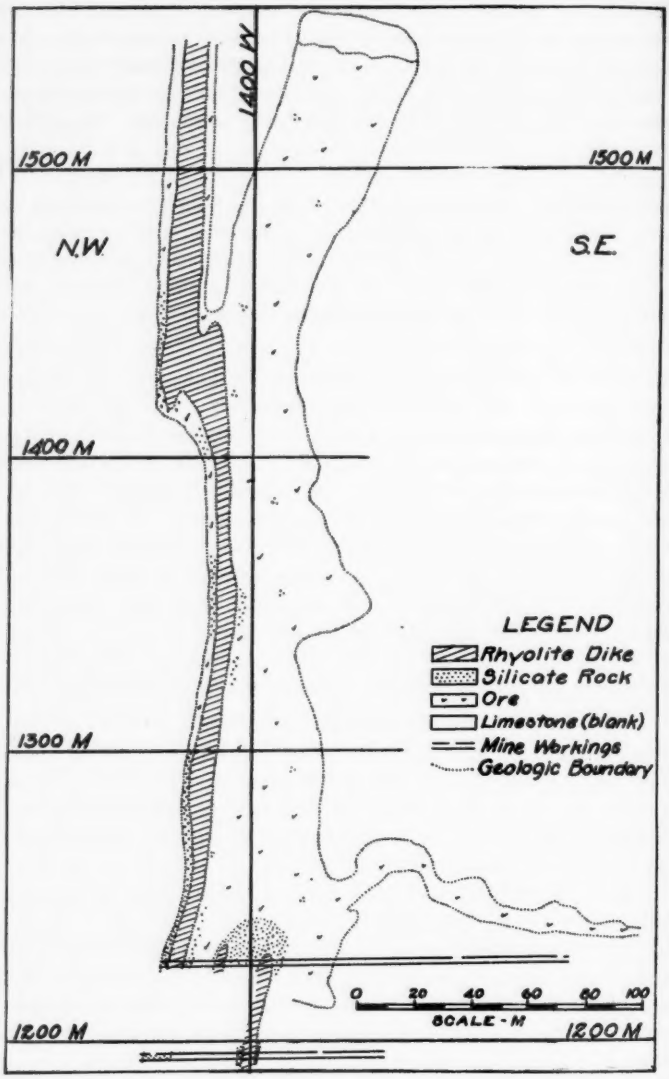


FIGURE 4.—*Tin Chimney*

Cross-section plane striking at right angles to trend of dikes, facing northeast

meager and is the low-grade silicate type. The middle dike is something new and is apparently a branch that has split from the eastern member. The ground between them, and for some distance above them, is intensely mineralized with garnet and allied silicates. Like all other silicates associated with dike contacts they are not only low-grade but are low in tin which is in distinct contrast to the overlying ore body. Their tin content does not average 0.10 per cent.

In contrast the tin content of the chimney proper might have been 1.5 to 2.0 per cent, which consequently raises a question as to its source. Its lowest known segment, only 10 meters thick, consists of unmined sulphide ore lying against the east wall of the eastern dike. These sulphides, underlain by limestone and grading upward into mined-out oxide, are the ones discussed under Field Observations, and their tenor is comparable to that of the Tin chimney's oxide material both in tin and remaining metallic content. Thus, either the ores are derived from the underlying silicate mass, or they come from a different source and have spewed into and around those silicates so as to suggest a silicate origin. If the latter is the correct solution the feeder may exist as a mantolike body at the elevation of the 10-meter sulphide segment.

There are occasional fragmentary remnants of the original oxides, and in addition records are available of the type of ore that was mined and milled. The ores did not differ physically from those expected in limestone camps. They were a drab, heterogeneous mixture of iron oxides with all the expectable variations ranging from beds of pure granular cerussite to areas dominantly zincy. Indeed, there was a 30-meter stretch from the 1300 to 1330-meter elevation that was entirely zinc ore. Vanadium minerals were unrecorded and have not been found in the fragmentary remnants. In addition, occasional areas carry relict structures of lime-iron silicates, but these silicate patches are a minor part of the ore and never approach the strength exhibited by the silicate cushion on which the chimney sits.

The cross section of Figure 4 shows an elliptical cap at the top of the chimney, an open cave according to old notes. This is a common feature in oxide ore bodies of limestone districts, whether they be mantos or chimneys, and is commonly referred to as a shrinkage cave. The origin of these caves is debated, but they are the result of oxidation. The sulphide bodies in Santa Eulalia show no such feature, and the ores are frozen cleanly to their limestone walls on all sides. Even the manto near the bottom of the chimney was accompanied by such a shrinkage.

The manto that leaves the chimney at an elevation of 1260 meters rises for a short distance and then loses elevation, thus giving the impression of having flowed down hill, which is contrary to the

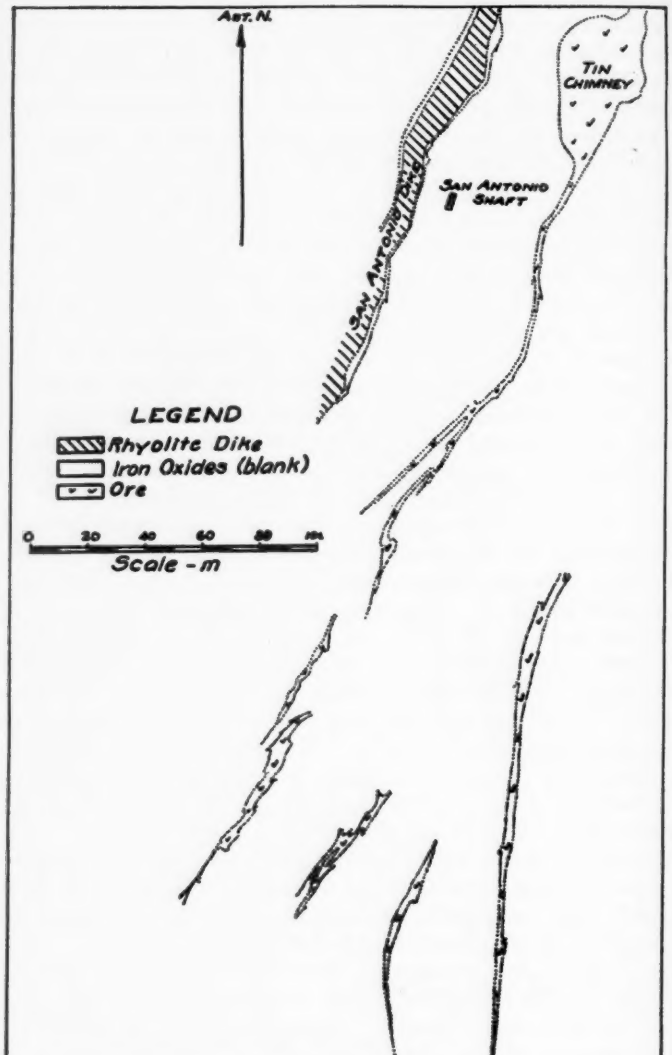


FIGURE 5.—*Dolores fissures*
Plan of part of the San Antonio 1st level

general belief concerning the emplacement of these bodies. It obviously is fed from the Tin chimney, and there is no other recognized source from which it could be derived. Like the chimney, it carried ores that averaged better than 1.0 per cent tin.

The Tin chimney passed through limestone into the conglomerates of the Capping series without any change in the character of the ore. It seems likely that it passed through the entire thickness of the Lower conglomerate and was terminated by one of the rhyolite sills of the Lower tuff. Oddly enough the top of the chimney is very close to the present erosion surface, lying within 60 meters of the hillside, yet surface showings give no clue to the proximity of the ore. The sole indication of underlying mineralization is one of the Dolores fissures which reaches the surface some 50 meters farther east as a narrow but strongly mineralized vein.

DOLORES FISSURES

From just below the 1500-meter elevation in the Tin chimney, a series of veins emerge from its eastern wall and rise flatly toward the east. These are the Dolores fissures (Fig. 5; Pl. 1). They vary in dip from 45° to 60°. Their average strike is somewhat west of south but in general is more southerly than the San Antonio dike. Near the chimney they are relatively close together, but southward they rapidly diverge, and in so doing some die and are replaced by new ones. They range from knife-blade streaks to shoots 10 meters wide, with an average commercial width of from 0.5 to 1.5 meters. The strongest of the group, the Yellow fissure, has been stoped for approximately 300 meters. Others have had commercial lengths of less than 40 meters.

These fissures are mineralized faults whose maximum throws, although not accurately measured, probably average 5-10 meters and rarely exceed 20. They cut the entire Capping series and extend downward into the underlying limestone where they are lost. Their greatest productivity has been from the Lower conglomerate, but they have produced ore from the Limestone series.

Their tin ores were of the same general grade as those of the Tin chimney, the tin being uniformly distributed throughout the entire group. Their mineralization is the customary mixture of lead and iron oxidation products common to all limestone bodies. However, in contrast to all other San Antonio bodies, neither fluorite nor any suggestion of a lime-iron silicate has yet been found, although feldspar has been reported from the single thin section prepared.

Thus, since the Dolores fissures show no signs of a lime-iron silicate phase, it can be argued that they represent a mineralization distinct from anything known within the mine; or, that the absence of silicates

indicates the fissures were not formed simultaneously with the emplacement of the silicate minerals, but possibly during or near the end of the sulphide stage and that they are related to other mineralizations within the mine; or, finally, that their absence is not indicative of either a distinct or later origin, but rather a change in mineralogy accompanying a change of temperature and depth. Elsewhere these silicates are found in limestone at elevations equivalent to the lower ranges of the Dolores fissures, and for that reason the final argument seems vitiated unless the possibility of lateral introduction from a distant source is considered. Although the writer favors the second suggestion a striking piece of evidence against it is that these are the only bodies apparently free of silicates.

In the Lower conglomerate ores exhibit an unusual cavernous structure made up of hollows with rounded to subangular walls set in a fine-grained, somewhat porous and siliceous matrix. The ore occurs as iron and lead oxidation products within the hollows and sometimes within the matrix and resembles a conglomerate from which the pebbles have been removed (Pl. 2, figs. 1-2). Erratic sulphides, about 12 meters below the first level at an elevation of approximately 1466 meters, show what were once limestone pebbles surrounded by a tuffaceous matrix, but now converted into pyrite, minor sphalerite, galena, and chalcopyrite (Pl. 2, figs. 3-4). A thin section from the same occurrence shows cassiterite grains shot indiscriminately through the sulphides and also what are probably orthoclase feldspars in the matrix.

The main source of impregnation for the fissures is thought to be the Tin chimney, and yet the fissures carry neither fluorite nor silicates. Moreover, occasional ore shoots of genuine importance, but isolated from the chimney by over 200 meters of apparently barren limestone, have shown no inclination to travel toward the chimney nor to connect with other veins having a chimney source. Therefore, the ore may have come, in part, from small bodies which sprang away from the dike contacts and then traveled laterally until they found and ascended these fissures, or the underlying sources may be associated with a blind ore body or an undiscovered contact between limestone and sill.

Since these fissures cut the Capping series their mineralizing solutions had to traverse all the associated rock types. The strongest veins pinch to knifelike streaks upon entering unfavorable beds and generally regain their strength upon emerging. A few fissures, however, have not regained their strength, while the mineralization of others upon hitting the upper surface of an unfavorable bed has spread along the contact as a narrow but widespread blanket. Consequently the mineralizers must have traveled laterally and vertically from different loci, for it is difficult to

conceive of a large fissure spreading widely upon reaching the top surface of an unfavorable horizon and not spreading equally widely along the bottom unless a lateral source had been feeding along the top of the bed. Although a few fissures are represented at the surface by strongly

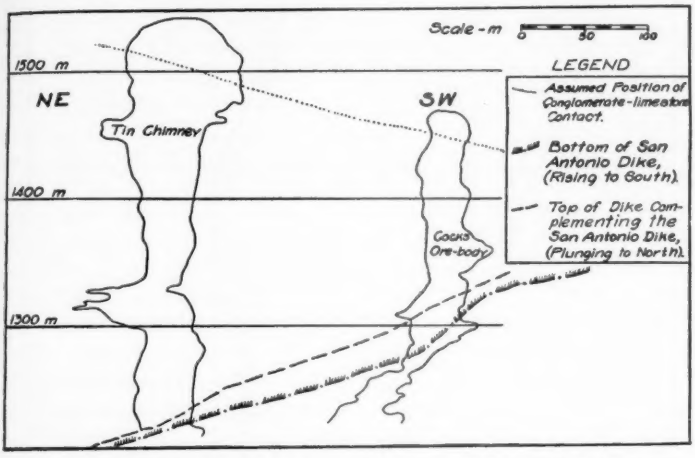


FIGURE 6.—*Longitudinal projection striking parallel to trend of dikes*
Facing S. 70° E.

mineralized but narrow veins, there is no commercial ore above the Lower conglomerate.

COCKS ORE BODY

Approximately 150 meters southwest of the Tin chimney, and on the opposite side of the dikes from it, the Cocks ore body forms a large and important chimney. Figure 6, a longitudinal projection along the dikes, shows the relationship between these two bodies, while Figure 7, a cross-sectional projection, shows the relationship between the Cocks ore body and the rhyolite dikes.

From its lowest level, or at about the 1225-meter elevation, this chimney inclines upward to the southwest, angles away from the dikes, away from the Tin chimney, and toward the West fault; and although it closely approaches the plane of the West fault, at an elevation of about 1300 meters, it never quite makes contact. Up to this point it has been a continuous although irregular ore body, swelling from the narrowest of mineable channels into large chambers of genuine importance. Where it closely approaches the West fault it turns abruptly and rises vertically with a large and fairly uniform cross section.

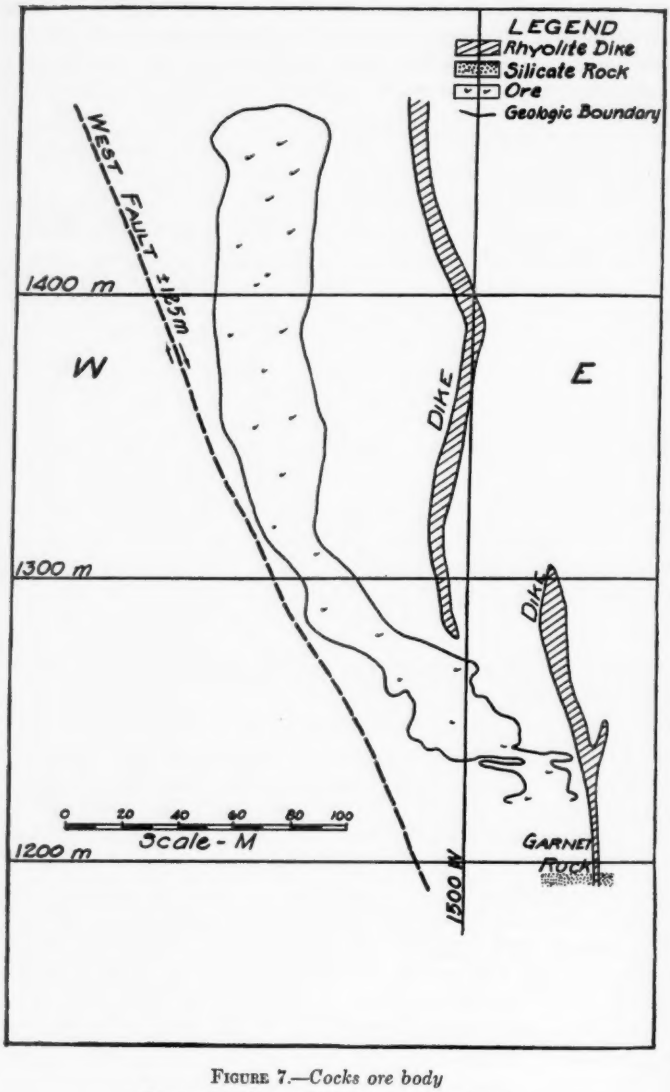


FIGURE 7.—*Cocks ore body*
Cross-section plane striking west-east, facing north

Like the Tin chimney, it is associated at its lowest elevation with the San Antonio dike, but where its projected position has been crosscut on the level below there is no ore body. Instead its position is taken by a mass of garnet and associated lime-iron silicates (Fig. 7). Also, it passes through the limestones into the Lower conglomerate and terminates, probably because of unfavorable tuffaceous or rhyolitic beds.

Its mineralogy varies throughout its length, and these variations seem to be primary, though accentuated by oxidation. It is associated with lime-iron silicates that form a shell surrounding and adhering to the fringes of the ore. At its deepest horizon, at 1225 meters, it is encased by garnet, while 30 meters higher the shell is less pronounced, and the garnet gives way to a mixture of garnet, tremolite, and coarse-fibered radiating silicates of actinolitic and hedenbergitic appearance. The garnets occur nearest the limestone and are followed inward by garnet and tremolite, or by a mixture dominantly hedenbergite or actinolite with minor epidote and garnet.

At an elevation of approximately 1285 meters the shell is confined mostly to the upper surface of this inclined section and consists of coarse radiating fibers of what seem to be hedenbergite. Above this the silicate shell becomes erratic, both in location and mineralogy. At 1315 meters silicates are sparse or absent. Their place is taken by a silicified jaspery rock that might be primary or secondary, while the few silicates found may be related to another ore body. At 1345 meters silicates reappear on one side as a strongly developed mass of garnet with fine to medium-fibered silicates, while 40 meters higher these give way to fibrous silicates that are distributed throughout the ore. Associated with them are silicification effects which again might be secondary. In the next 40 meters all signs of either silicification effects or silicates disappear.

Likewise the metallic content varies, and the body is divided into three distinctive zones. From the top down, we find a silver-lead-tin zone, a lead-vanadium zone, and finally a silver-lead zone overlying primary mineralization. No sulphides have been developed.

The silver-lead-tin zone extends from the top down about 130 meters and terminates at about the 1345-meter elevation. It is a mass of iron oxides liberally sprinkled with lead and zinc minerals and does not differ from the tin ores of the Tin chimney; its tin content was only slightly lower.

Below 1345 meters the tin content decreased, while between 1345 and 1280 meters the ore body changed markedly in both mineralogy and internal structure. This zone, marked by lead-vanadium ores, lies directly at the bottom of the vertical section of the chimney and close to the West fault.

Below the lead-vanadium zone the chimney regains a more normal metallic content, resumes a normal internal structure, but loses its well-defined vertical character. All vanadium content vanishes, and although a meager amount of tin is present, occasionally as high as 0.50 per cent, it is too low to be mined for anything but silver-lead ore. It is this section that becomes highly irregular, although continuous, and leads downward to the northeast. It carries no special problems as does the lead-vanadium zone.

Between 1280 and 1285 meters the ore changes abruptly from cerussite sand to an overlying bed of vanadinite sand 2.0-2.5 meters thick, which in turn passes upward into a zone of stratified manganese muds with intercalated layers of powdery vanadinite. Above these there is about 40 meters in which the entire ore body is completely fragmental and in which the fragments have been rehealed by a vanadium-bearing cement. At the bottom of the zone the particles are healed completely by lead-vanadium salts, but at the top the fragments are merely coated with vanadium minerals, and the body of the cement is calcite (Pl. 2, fig. 5).

The fragments range from pea size to 2-meter boulders and include iron oxides, slabs of galena, cubes of fluorite, hard white radiating silicates, argentiferous cuprite, limestone, and possibly rhyolite. This part of the ore body passes upward into a solid unbroken ore in which vanadium salts are absent or spotty and low grade.

The entire mine has been undergoing extensive post-ore solution. Caves are numerous, especially at the vanadium-bearing horizon, and most of them lie with their long axes horizontal. They range from small chambers to rooms 300 meters long and 30 meters wide. Following the emplacement of the sulphides, and during or following their oxidation, a cave might develop just beneath and possibly within the ore body where it turns upward into a vertical chimney. Following this the picture calls for movement along the West fault, collapse of overhanging material into the underlying cave, and cementation by secondary vanadium minerals.

Support for this idea is found at the bottom of the vanadium zone where the vanadinite sands and the manganese muds have been deposited in a cave. This cave, open at the time of discovery, had a limestone back. Furthermore, just below the middle of the vanadium zone another long horizontal cave angles gently toward the West fault and enters the fragmental zone. The entrance to this cave shows a definite transition from fragmental ore body to a comparatively open cave with the entrance obscured and completely choked by calcite and gypsum-blanketed rubble. In parts of the mine distant from the West fault, caves have penetrated ore bodies, but the caves have remained open, have not been choked by

fragmental ore, and only under special circumstances are they filled with limestone rubble.

The source of the vanadium is a separate problem.

VANADIUM ORES

In contrast to the tin ores, commercial vanadium in San Antonio is a function of secondary concentration and reflects the eccentricities of oxidation and migration. The ores have come from the breakdown of primary vanadium-bearing minerals, the leaching of vanadium salts from these minerals, their attendant transportation in solution, and finally their precipitation in their present form. The problem is not the origin of the vanadium ore minerals but rather the form of the primary vanadium. It seems likely the ores have come in part from the oxidation of vanadium-bearing silicates and in part from vanadium-bearing sulphides. Field evidence suggests that their major source is the vanadium-bearing silicates, but no primary vanadium-bearing mineral has been identified.

The ores lack distinctive minerals and features. They range from yellowish coatings surrounding all types of rock fragments to stratified manganese muds in which yellowish vanadium-bearing streaks are intercalated between layers of wad or zones of gypsum. Sometimes the wad and gypsum carry vanadium. Most of the vanadium-bearing material powders readily and when wet breaks into slimes.

They also occur in stratified, slightly consolidated beds of darkish-yellow, sandy minerals. Although granular, the individual grains are too small to be distinctive. The beds range from a few centimeters to 2.5 meters thick, with each bed showing a fairly constant thickness. They are confined within the walls of once-open caves and are neither widespread nor abundant.

Along the backs and walls of some caves calcite-coated vanadinite stalactites range from incipient bumps to pieces 20 centimeters long and 3 centimeters in diameter. They are hard, in contrast to the powdery types, and show a feathery internal structure. Dominantly they are one mineral, but toward the edges they show alternating rings of vanadinite and calcite. Their color suggests that these are vanadium-bearing.

Judging from mine assays most of the ore is vanadinite with some descloizite and mimetite. These minerals are represented in all the types except the stalactitic which seems to be composed entirely of vanadinite. Attempts to find additional vanadium minerals, such as the calcium vanadates, were without success. All these types are products precipitated in pre-existing openings—limestone caves, caves developed within oxidized ore bodies, and the porous and oxidized mass of the ore body itself.

The ores occupy a definite horizon about 60 meters thick, whose bot-

tom corresponds with the bottom of the lowest cave horizon in the down-faulted block. Not all of this 60-meter zone produced vanadium, and certain large areas, notably the region adjoining the Tin chimney, are nearly barren. On the other hand those areas whose upper levels contain an extensive development of oxidized fibrous silicates are the areas that produced vanadium ore bodies at a lower horizon, albeit there is sometimes a considerable spread between the location of the overlying silicates and the underlying vanadium concentration. Consequently a possible correlation can be made between overlying oxidized silicates and underlying vanadium ore.

Newhouse (1934) concluded that vanadium ores, associated with the oxidized zones of lead, and lead-zinc-copper deposits are derived from small amounts of vanadium present in the original sulphides. Although sulphide ores have not been developed in the San Antonio district there are occasional exposures of small sulphide bodies, and from these exposures sulphide samples have been assayed for vanadium. The majority showed a trace or nothing, but samples from one section did show the following vanadium percentages: 0.32, 0.32, 0.27, and 0.27. It is likely therefore that primary sulphides furnished part of the vanadium.

However, assaying bears out the possibility, suggested by field evidence, of a correlation between silicates and vanadium ores. In only a few of the numerous unoxidized silicate samples assayed was vanadium absent. Although many ran but a trace or better, a small percentage showed 0.23 per cent or higher, while one ran 0.50 per cent. Four assays not confined to any one section of the mine, show the following: 0.23, 0.27, 0.27, and 0.50 per cent.

Samples composed dominantly of garnet produced the lowest figures; those with fibrous silicates contained more vanadium; but samples with fine- to medium-textured, light-colored fibrous silicates consistently held most vanadium. Whether vanadium replaces iron molecules in otherwise normal lime-iron silicates or occurs as a specific vanadium silicate is not known. However, the light-colored, fine- to medium-textured fibrous silicates are regarded as the vanadium carrier, although other silicate minerals may contain some vanadium.

OXIDIZED ZINC ORES

Zinc ores are found in approximately the same zone as the vanadium but extend deeper. Areally, however, the oxidized zinc and the vanadium occupy distinctly different locations, and the two have not been found together. In contrast to vanadium ore bodies, the zinc ores are closely associated with oxidized iron and lead ore bodies and underlie areas in which oxidized silicates are not pre-eminent. They are rarely found as

precipitates on pre-existing surfaces but seem to be replacements in limestones or iron-oxide mineralizations. They carry hemimorphite (calamine) and smithsonite and numerous earthy mixtures of iron and zinc.

ORIGIN

Since the San Antonio deposits are metasomatic replacements of limestone by invading metasomes, the question arises as to the nature of the invading metasomes and the conditions under which they operated. The mineral assemblages suggest high temperature, yet stratigraphic and physiographic evidence indicates that the deposits formed at no excessive depth. These deposits can probably be placed in one of the standard groupings, and the classification most suited to them can be chosen from among the contact-metamorphic, hypothermal, pyrometasomatic, and pneumatolytic types.

If these are not contact-metamorphic deposits dependent upon the known intrusive they might still be of contact origin but related to an unexposed intrusive. If so, they are best placed in Lindgren's pyrometasomatic class. As for their being of hypothermal origin, Lindgren (1933, p. 637) includes among the indicative minerals "pyroxenes and amphiboles, the garnets, ilmenite, magnetite, specularite, pyrrhotite, tourmaline, topaz, . . ." and further states that "simple sulphides and arsenides prevail and are in many deposits associated with oxides such as magnetite, specularite, ilmenite, and cassiterite." These minerals would seem to fit San Antonio. In Lindgren's (1933, p. 698-701) discussion of pyrometasomatic deposits it becomes extremely difficult to distinguish between his hypothermal and pyrometasomatic groupings, except by the absence or presence of an igneous contact. In this case a limestone deposit is under discussion, and it is not desired to choose between the two types on such a basis. Limestones are so permeable to invading mineralizers and regardless of the position of a contact allow mineralization to be accomplished with such ease as to invalidate the distinction.

The following are the known primary minerals in the San Antonio deposits, those marked with an asterisk not being positively identified.

pyrite	quartz	idocrase*
chalcopyrite	cassiterite	ilvaite
pyrrhotite	wolframite	scapolite
sphalerite	columbite	topaz
arsenopyrite	garnet	tourmaline
galena	epidote	orthoclase
magnetite	hedenbergite*	muscovite
ilmenite	tremolite	fluorite
hematite	actinolite	

In addition there is a group of unknown but recognized minerals.

Recalling the mineral assemblage accompanying cassiterite, discussed under Petrographic Observations, and referring to pneumatolytic as de-

fined by C. N. Fenner (1933, p. 71),—"pertaining to processes or results effected by gases evolved from igneous magmas or entrained with gases of that origin,"—there is a possibility the deposits may be partly pneumatolytic. At the same time they show all the characteristics of contact metamorphism and probably belong in the pyrometasomatic group.

SUMMARY OF CONCLUSIONS

(1) The chimney-shaped replacement deposits of San Antonio lie in a thick series of Cretaceous limestones. Fossil suites suggest ages ranging from Glen Rose to Upper Cretaceous.

(2) The deposits are closely associated with a zone of acid dikes that lie within a downfaulted block. The dikes, rhyolites at the surface but felsitic in depth, controlled the location of the ore bodies but were not the source of the mineralizers.

(3) Cassiterite introduction, judging from field and petrographic evidence, followed the emplacement of lead, iron, and zinc sulphides, although there may have been an uninterrupted sequence of mineralization.

(4) Field evidence suggests that the vanadium ores were formed from the reorganization of primary vanadium-bearing silicates as well as vanadium-bearing sulphides.

(5) Although the intrusive responsible for the mineralization has not been found, the deposits are placed in the pyrometasomatic group and in part may have been formed by pneumatolytic action.

REFERENCES CITED

Fenner, C. N. (1933) *Pneumatolytic processes in the formation of minerals and ores,* in, *Ore deposits of the Western States,* Lindgren volume, p. 58-106.

Kelly, W. A. (1936) *Evolution of the Coahuila peninsula, Mexico, Part II. Geology of the mountains bordering the valleys of Acatita and Las Delicias,* Geol. Soc. Am., Bull., vol. 47, p. 1009-1038.

Kellum, L. B., Imlay, R. W., and Kane, W. G. (1936) *Evolution of the Coahuila peninsula, Mexico, Part I. Relation of structure, stratigraphy, and igneous activity to an early continental margin,* Geol. Soc. Am., Bull., vol. 47, p. 969-1008.

Lindgren, Waldemar (1928) *Mineral deposits,* 3d ed., p. 796; rev. ed. (1933) p. 637, 698-701.

Newhouse, W. H. (1934) *The source of vanadium, molybdenum, tungsten, chromium, in lead deposits,* Am. Mineral., vol. 19, p. 209-220.

Prescott, Basil (1916) *The main mineral zone of the Santa Eulalia district,* Am. Inst. Min. Metall. Eng., Tr., vol. 51, p. 57-99.

——— (1926) *The underlying principles of the limestone replacement deposits of the Mexican province,* Eng. & Min. Jour., vol. 122, p. 246-253, 289-296.

American Smelting and Refining Co., Apdo. 3, Bis., Chihuahua, Chihuahua, Mexico.
Manuscript received by the Secretary of the Society, April 15, 1941.

EXPLANATION OF PLATE 2

PHOTOGRAPHS OF SAN ANTONIO ORES

Figure

(1) Oxide ores from the Dolores fissures illustrating cavernous structure typical in the Lower conglomerate.

(2) Detail from Figure 1 showing rough irregular oxidation products of lead and iron within the pebble casts.

(3) Polished plate, under reflected light, of erratic sulphides found in the Dolores fissures at the horizon of the Lower conglomerate. It shows limestone pebbles selectively replaced by sulphides.

(4) Polished plate of Figure 3 under direct light. The pebblelike bodies are iron sulphide with minor amounts of lead and zinc. The surrounding tuffaceous matrix shows only slight replacement.

(5) Typical fragmental ore from the vanadium horizon in the Cocks ore body. The dark fragments are iron oxide coated with a light-colored band of vanadinite. The crinkly botryoidal material above number 4 is also vanadinite, while the white patch and the colorless crystals above number 0 are calcite.

Figure 1

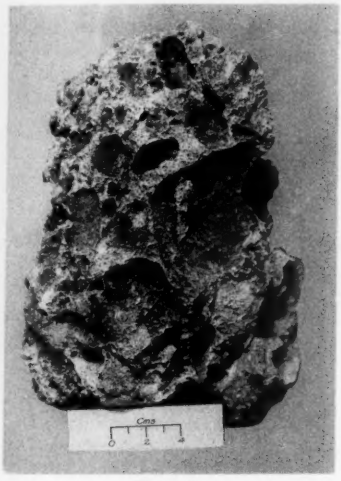


Figure 2

Figure 3

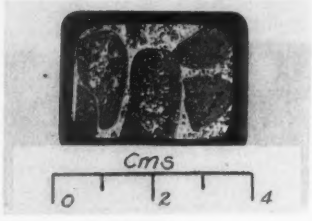


Figure 4

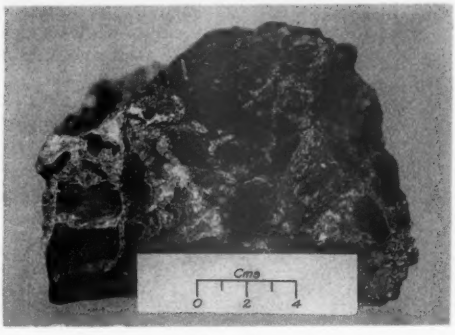


Figure 5

PHOTOGRAPHS OF SAN ANTONIO ORES

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 54, PP. 205-262, 5 PLS., 3 FIGS. FEBRUARY 1, 1943

STRATIGRAPHY AND STRUCTURE OF SOUTHWEST SAYPO QUADRANGLE, MONTANA

BY CHARLES DEISS

CONTENTS

Page

Abstract 207
Introduction 207
Location of area 207
Acknowledgments 209
Summary of previous work in area 209
Belt series 211
General statement 211
Miller Peak argillite 211
Cayuse limestone 215
General statement 215
Original description 216
Hoadley formation 216
Original description 216
Ahorn quartzite 217
Original description 217
Cambrian system 218
General statement 218
Flathead sandstone 219
Gordon shale 219
Damnation limestone 220
Dearborn limestone 220
Pagoda limestone 221
Pentagon shale 221
Steamboat limestone 222
Switchback shale 223
Devils Glen dolomite 223
General considerations 223
Cambrian section on Allan Mountain 223
Devonian system 224
General statement 224
Devonian section on Allan Mountain 226
Carboniferous system 227
General statement 227
Hannan limestone 228
General considerations 228
Original description 229
Jurassic system 231
General statement 231

Page

Age of Ellis formation in Saypo quadrangle 231
Jurassic section at head of Lime Gulch 233
Cretaceous system 236
Kootenai formation 236
Colorado shale 238
General statement 238
Lithology of Colorado shale in Sawtooth Range 239
Quaternary system 240
General statement 240
Pleistocene deposits 240
Recent alluvium 240
Hiatuses within map area 241
General statement 241
Cambrian—pre-Cambrian unconformity 241
Devonian—Cambrian hiatus 242
Mississippian—Devonian hiatus 243
Jurassic—Mississippian hiatus 243
Tertiary erosion 244
Igneous rocks 245
Type and occurrence 245
Composition of sills 247
Deformation of sills 248
Structural provinces in southwest Saypo quadrangle 248
Faults 249
General statement 249
High-angle thrust faults 249
Low-angle overthrust faults 250
Low-angle normal fault 251
High-angle normal faults 251
Folds 252
General statement 252
Eastern belt of folding 252
Western belt of folding 252
Sequence of geologic events 253
Proterozoic to Cenozoic sequence 253
Sequence during Laramide orogeny 255
General statement 255
Early compressional phase 255
Late tensional phase 257
Post-Laramide sequence 257
Conclusions 258
References cited 260

ILLUSTRATIONS

Figure Page

1. Map showing location of (1) Saypo, (2) Silvertip, (3) Ovando and (4) Coopers Lake quadrangles 208
2. Index map of area 214
3. Structure sections across southwest part of Saypo quadrangle 246

Plate — Facing page
1. Hannan limestone ... 230
2. Cirques and perched erratic ... 231
3. High-angle thrusts ... 248
4. Folded diorite sills ... 249
5. Normal and overthrust faults ... 252

ABSTRACT

The Saypo quadrangle, in northwestern Montana, lies between parallels 47°30′ and 48° and meridians 112°30′ and 113° and contains parts of the Lewis and Clark and the Sawtooth ranges and part of the northern Great Plains. Stebinger mapped the geology west to the mountains before 1918, and the writer began mapping the mountains in 1940 for the United States Geological Survey.

The stratigraphic column, approximately 14,500 feet thick, consists of 19 formations: four of Belt, nine of Cambrian, two of Devonian, and one each of Mississippian, Jurassic, and Lower and Upper Cretaceous age. Pleistocene drift and Recent alluvium lie on all older rocks. Belt and Cretaceous sediments contain gabbro and diorite sills. Unconformities occur between Cambrian and Belt, Devonian and Cambrian, Mississippian and Devonian, and Jurassic and Mississippian formations. None are strongly angular.

The region contains two structural provinces: an eastern province characterized by westward-dipping, high-angle thrust faults, which broke steeply dipping Paleozoic and Mesozoic sediments, and a western province characterized by westward-dipping, low-angle overthrusts and westward-dipping, high-angle normal faults which broke the Belt and older Paleozoic sediments. Folds are subordinate to faults, both in number and importance.

All major structures probably formed during the Laramide orogeny. Its earlier compressional phase produced open folding, then high-angle thrusting, and finally low-angle overthrusting. The later tensional phase resulted in low-angle normal and, toward its close, in high-angle normal faulting.

INTRODUCTION

LOCATION OF AREA

The eastern front of the Rocky Mountains of northwestern Montana is represented within the Saypo quadrangle by the Sawtooth Range which is bounded on the west by the valleys of the North and South Forks of Sun River, and in turn by the Lewis and Clark Range. The Continental Divide lies just west of and crosses the northwest corner of the quadrangle (Fig. 2).

The Saypo quadrangle is bounded by the parallels 47°30′ and 48° N. Lat. and by the meridians 112°30′ and 113° W. Long. (Fig. 1). The southwest part of Teton County, part of the north portion of Lewis and Clark County, and the eastern part of Flathead County are within the quadrangle. The western two-thirds of the quadrangle is largely within the Lewis and Clark National Forest. The northwest part is in the Flathead National Forest. The east front of the Sawtooth Range is approximately 65 miles west of Great Falls and 23 miles west of Choteau. The quadrangle is covered by a topographic map on the scale 1/125,000, and with the contour interval of 100 feet.

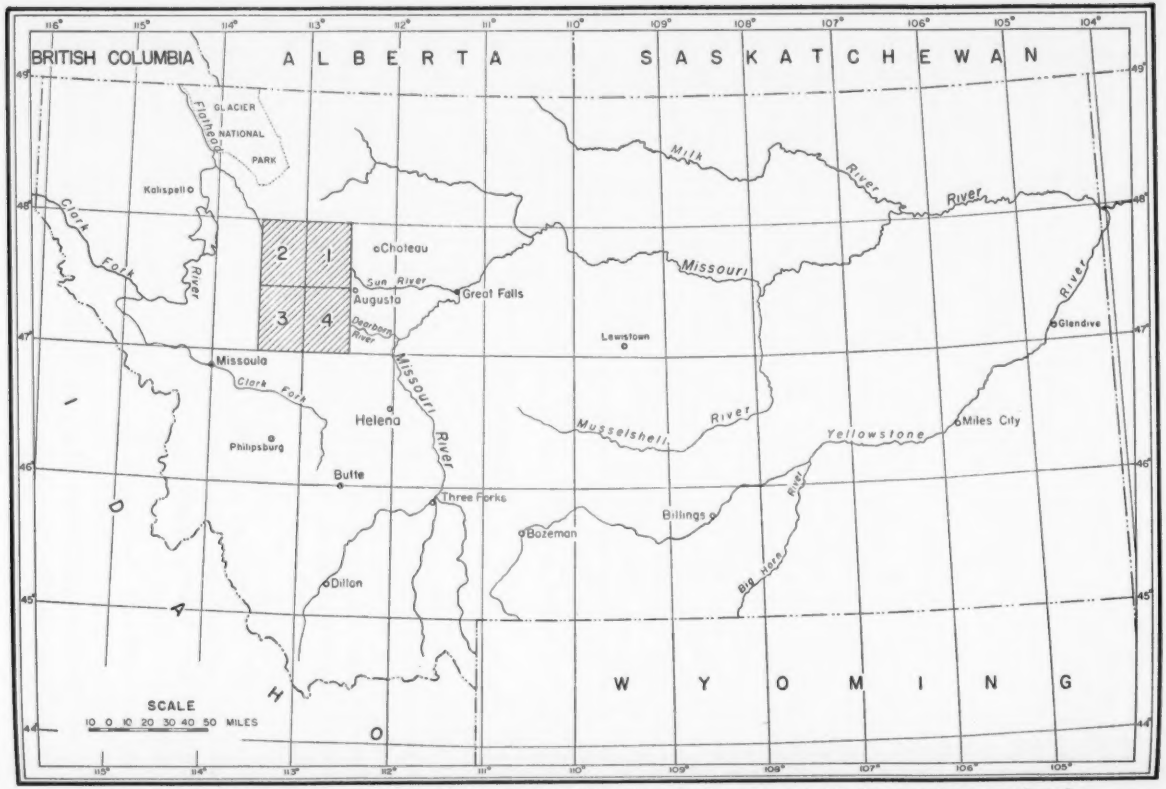


FIGURE 1.—*Map showing location of (1) Saypo, (2) Silvertip, (3) Ovando and (4) Coopers Lake quadrangles*

The geographic names used in this paper are in current local use and are published on the official maps of the Lewis and Clark National Forest by the United States Forest Service. Many of the names of streams and mountain peaks are different from those given on the Saypo quadrangle which was surveyed in 1901 and published in 1903. (*Compare* Figure 2 and a copy of the Saypo quadrangle.)

ACKNOWLEDGMENTS

Drs. Josiah Bridge and G. R. Mansfield read the manuscript and gave many helpful suggestions. The fossils collected from the Cretaceous and Jurassic rocks were identified by Dr. John B. Reeside, Jr. who furnished the information concerning their age and correlation and also read the part of the manuscript dealing with the Mesozoic. Dr. G. A. Cooper identified most of the Devonian fossils, established their age, and read the discussion of the Devonian. Drs. R. S. Bassler, Edwin Kirk, and Roland W. Brown also identified some of the fossils. Dr. Clarence S. Ross identified the igneous rock specimens and furnished much of the information concerning their mineral content. The writer was ably assisted in the field by Dr. Robert M. Garrels whose interest and untiring energy added much to the mapping completed in 1940. To each of these men the writer is deeply indebted and extends his grateful appreciation and thanks.

SUMMARY OF PREVIOUS WORK IN AREA

The first report concerning the rocks and geologic structure of the Sawtooth and Lewis and Clark ranges was by Chapman (1900, p. 153-156) who published a reconnaissance section across the Rocky Mountains from the crest of the Mission Range, northeast across the Swan, Lewis and Clark, and Sawtooth ranges to the plains. Chapman first described the rocks of the Sawtooth and Lewis and Clark ranges and assigned them to the Cambrian and Carboniferous systems. Chapman interpreted all the faults which he recognized in the mountains as high-angle normal faults. He also noted the igneous rocks in the Saypo quadrangle and the associated hot springs.

Willis worked in the Lewis and Livingston ranges in Glacier National Park, 35 to 60 miles northwest of the Saypo area. His studies resulted in his discovery and description of the Lewis overthrust and the relationships of the Belt rocks in the mountains to the Cretaceous rocks on the plains (Willis, 1902, p. 305-349) and have important bearing on the structures in the mountains in the Saypo quadrangle. Willis' results suggested the possible extension of the Lewis overthrust southward and also indicated the age and lithologic characteristics of the Belt rocks in the front ranges of the Rockies in northwestern Montana.

FIGURE 1.—*Map showing location of (1) Saypo, (2) Silvertip, (3) Ovando and (4) Coopers Lake quadrangles*

The Saypo area was not studied geologically until 1913 when W. O. Crosby (Powers and Shimer, 1914, p. 556), in preparation for locating a dam, examined the Sun River Canyon between the mountain front and the junction of the North and South Forks of Sun River. Faunal lists from five fossil localities and Crosby's notes on the structure and rocks in the area were published by Powers and Shimer (1914, p. 557 and 559), who first noted that each parallel ridge in the Sawtooth Range marks the position of an overthrust fault to the east.

Two years later, Stebinger (1918, p. 150) did the first accurate geologic mapping in the Saypo quadrangle when he began his field work on the plains east of the Sawtooth Range. Stebinger's (1918, Pl. 24) map of the Birch Creek-Sun River area covers the eastern part of the quadrangle to the mountain front (Fig. 2) and will be combined with the map being made in the Sawtooth and Lewis and Clark ranges to complete the Saypo quadrangle.

Between 1918 and 1933 several geologists of the Montana Bureau of Mines and Geology made reconnaissance traverses across parts of the Saypo quadrangle. Bevan (1929, p. 446-449) described the structure of the Sawtooth Range (incorrectly referred to as the "Lewis and Clark Range") as a "Zone of overthrust slice blocks" and suggested that "the structure resembles the Scottish Highlands type." In 1931, 1932, and 1933, C. H. Clapp and the writer mapped most of the Coopers Lake quadrangle and carried the mapping into the south and west parts of the Saypo quadrangle. In connection with stratigraphic studies on the Paleozoic rocks, the writer did reconnaissance mapping in the west part of the quadrangle in 1932 and 1933.

C. H. Clapp (1932, Pl. 1) published a geologic map of northwestern Montana on the scale 1/500,000, showing the Belt rocks in the Saypo quadrangle as one formation, as are the combined Cambrian and Devonian, and the Kootenai, Colorado, and Montana units. The contacts of the formations and the geologic structures in the northern part of the Sawtooth Range were interpreted and compiled from the topography on the Saypo and Heart Butte quadrangles. Nevertheless, the map is reasonably accurate for the scale used and is the only geologic map yet published of the western two-thirds of the Saypo quadrangle.

The complex structure in the Coopers Lake quadrangle and the little-known stratigraphy of the Paleozoic formations prompted the writer, between 1931 and 1936, to study the stratigraphy and fossils of these rocks, particularly of the Cambrian in the Swan, Flathead, and Lewis and Clark ranges. This work resulted in a fairly mature grasp of the stratigraphic problems and in familiarity with the Paleozoic rocks likely to be encountered in geologic mapping of the Sawtooth and Lewis and

Clark ranges. The work also resulted in several publications as the studies progressed, but none of the papers dealt specifically with the geology of the Saypo quadrangle.

In 1940 mapping was done in the Lewis and Clark and Sawtooth ranges in the southern part of the Saypo quadrangle as part of the field work of the Section of Areal and Nonmetalliferous Geology, of the Geological Survey, United States Department of the Interior. The areal geology was mapped on the scale 1/62,500 by means of a 10-inch open sight alidade and a 15-inch plane table. The Saypo topographic map published in 1903 was enlarged photographically and formed the topographic base for the geologic mapping. The formations exposed in the southwestern part of Saypo quadrangle are shown in Table 1.

BELT SERIES

GENERAL STATEMENT

The oldest rocks exposed in the Rocky Mountains of northwestern Montana are Belt (pre-Cambrian) in age. In the Saypo quadrangle only the upper Belt rocks are exposed. They comprise much of the Lewis and Clark Range and most of the outcrops in the western part of the quadrangle but are not exposed in the Sawtooth Range east of the North and South Forks of Sun River (Fig. 2).

The Belt rocks in the Saypo quadrangle are quartzitic sandstones, argillites, and impure sideritic and limonitic limestones (Pl. 5, fig. 2). The rocks are divided into four formations which, from the base upward, are: (1) Miller Peak argillite, (2) Cayuse limestone, (3) Hoadley formation, and (4) Ahorn quartzite. These formations are equivalent in age to the lower and middle part of the Missoula group of Clapp and Deiss (1931, p. 677-683, Fig. 2) in the Sapphire and Garnet ranges of western Montana. Precise correlations of the formations in the two localities have not been worked out, but the Miller Peak argillite has been traced by mapping from its type section into the Saypo quadrangle.

MILLER PEAK ARGILLITE

The Miller Peak argillite originally was defined from its type section on Miller Peak (Clapp and Deiss, 1931, p. 678) in the Bonner quadrangle, Montana, approximately 70 miles southwest of the outcrops in the Saypo area (Fig. 1). From 1930 to 1933, in connection with stratigraphic studies on the Belt series and with mapping the Coopers Lake quadrangle, Clapp and Deiss (1931, p. 683-688) traced and mapped the Miller Peak and overlying formations of their Missoula group from the type section to the central part of the Lewis and Clark Range (Deiss, 1935, p. 96).

TABLE 1.—*Formations exposed in southwestern part of Saypo quadrangle*

System	Series	Formation	Character	Thickness in Feet
Quaternary	Recent	Alluvium	Sands, gravels, and muds generally restricted to larger valley floors where they form flood plains.	
	Pleistocene	Glacial drift	Undifferentiated moraines and fluvioglacial deposits. Possibly some outwash mapped as Pleistocene is Recent alluvium.	1–250?
		—Unconformity—		
Cretaceous	Upper Cretaceous	Colorado shale with Blackleaf sandy member at base	Upper part black fissile shale and few widely spaced thin sandstone beds. Lower part, Blackleaf sandy member, is alternating thick units of black shale and thin units of nodular-bedded sandstone. One bed contains marine pelecypods. In upper part of Blackleaf member is 60-foot zone of manganiferous shale.	1000?
	Lower Cretaceous	Kootenai formation	Tan and green-gray sandstone in lower part overlain by maroon and green fissile and chunky shale and bright-red soft sandstone which grades up into green-gray thick- and thin-bedded sandstone. In middle is 25- to 30-foot unit of chert-pebble conglomerate overlain by lenticular beds of fossiliferous limestone. Upper part of Kootenai tan-gray and pink hard banded cross-bedded quartzitic sandstone.	900+
Jurassic	Upper Jurassic	Ellis formation	Basal hard sandstone overlain by dark- and light-gray fissile and chunky shale interbedded with 1- to 23-foot units of pale-gray fine-grained argillaceous and arenaceous limestones. Marine fossils of Divesian and Argovian age abundant at four widely separated horizons. Upper part of formation sandy.	250
		—Unconformity—		
Carboniferous	Mississippian	Hannan limestone	Basal limestone-breccia overlain by gray limestone and alternating units of shaly-bedded argillaceous limestone which grades upward into hard tan- and pale-gray crystalline fossiliferous limestone; contains much pale-gray chert in nodules and beds as much as 7 inches thick. Upper 200 feet consists of white-gray finely crystalline thin- and thick-bedded dolomite.	1370
Devonian	Upper Devonian	Two new unnamed formations	Dolomite limestone and shale overlain by tan-gray finely to medium-crystalline fossiliferous limestone. Alternate units of drab-brown, petroliferous, arenaceous, thick-bedded dolomite and light-gray, hard, finely crystalline, thick- and thin-bedded limestone. All more or less petroliferous and argillaceous.	980
		—Unconformity—		
	Upper Cambrian	Devils Glen dolomite	White-gray to pale-buff-gray, finely crystalline, thick- and some thin-bedded dolomite. Mottled salmon-pink and coarser-grained in upper part of formation. Weathers white gray and forms sheer cliffs.	198–250
	—?—	Switchback shale	Green and gray soft shale, locally calcareous and arenaceous, interbedded with flaggy-bedded magnesian limestone which weathers rusty tan.	75–125
		Steamboat limestone	Pale-gray and tan fine-grained thick- and thin-bedded limestone overlain by interbedded tan-gray irregularly bedded fossiliferous limestone and green fissile shale. Shale in units 2 to 8 feet thick separated by much thicker units of limestone.	275

TABLE 1.—*Continued*

System	Series	Formation	Character	Thickness in Feet
Cambrian	Middle Cambrian	Pentagon shale	Buff-gray platy calcareous thick-bedded extremely fossiliferous shale which grades upward into thin- and thick-bedded limestone.	0–290
		Pagoda limestone	Green fissile shale interbedded with thin-bedded limestone, overlain by cream- and chocolate-gray finely crystalline limestone, thin-bedded in lower and upper parts, and oolitic and thick-bedded in middle part.	300
		Dearborn limestone	Green fissile shale and intercalated limestone conglomerate, overlain by tan-gray thick- and thin-bedded limestone which contains varying amounts of buff and orange-tan clay as flakes, nodules and partings.	300
		Damnation limestone	Blue- and tan-gray thin-bedded fossiliferous limestone which contains much limonitic clay, overlain by chocolate-gray and tan fine-grained thick- and thin-bedded hard massive limestone which also contains flakes and blebs of buff siliceous clay.	150
		Gordon shale	Drab-green-gray and maroon fissile to chunky micaceous shale interbedded in upper-middle part with thin beds of sandstone and thicker beds of limestone. Upper beds contain many fossils.	220
		Flathead sandstone	Coarse- and fine-grained cross-bedded sandstone interbedded with thin units of micaceous and arenaceous shale. Most diagnostic characteristic is presence of white quartz pebbles ⅛ to 3 inches in diameter. Basal beds usually pebbly conglomerate.	50–100
		—Unconformity—		
pre-Cambrian	Belt	Ahorn quartzite	Pink thick-bedded medium- to coarse-grained arenaceous and pure quartzite and occasional beds of red sandstone in lower 1700 feet, overlain by green and red-gray thin-bedded argillite and a few intercalated thin beds of fine-grained sandstone in upper 400 feet.	2100
		Hoadley formation	Dark-green-gray and maroon, thick-bedded, siliceous argillite and some vitreous, fine-grained quartzite overlain by red and buff, arenaceous, calcareous argillite and fine-grained sandstone which grades upward into greenish-buff-weathering, calcareous sandstone and much interbedded pink sandstone and argillite. Upper part brilliant- and dark-red soft thin- and thick-bedded platy ripple-marked sandstone interbedded with some pale-red argillite.	4100
		Cayuse limestone	Dull-gray, crystalline, buff-tan-weathering dolomite interbedded with pale-maroon and green-gray, fissile argillite which grades upward into alternating zones of shaly bedded, calcareous argillite, and finally passes into sideritic and dolomitic hard marble. Upper part blue- and dull-gray, thick-bedded, finely oölitic, and coarsely crystalline algal limestone which shows typical bioherm structure.	1000
		Miller Peak argillite	Pale-green and maroon interbedded, thin- to fissile-bedded, hard, siliceous argillite overlain by ripple-marked argillite and intercalated dolomitic and quartzitic sandstone, which grade upward and become more dolomitic passing into the overlying Cayuse limestone. Base cut out by Lewis overthrust.	1000 exposed

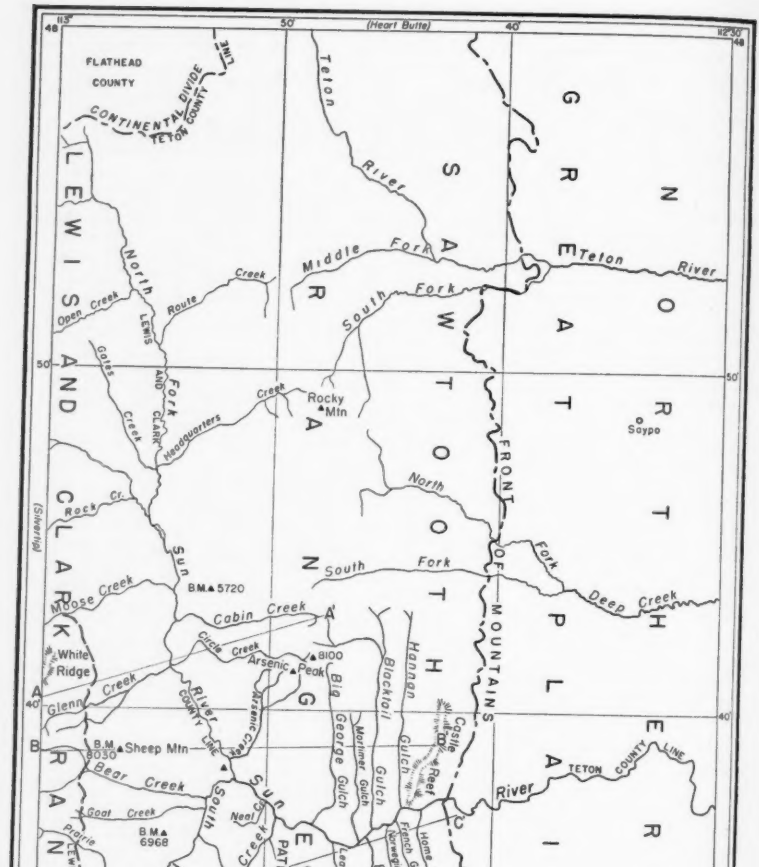


FIGURE 2.—*Index map of area*

The Miller Peak argillite is exposed just west of the South Fork of Sun River throughout a distance of several miles north and south of Deer Creek (Fig. 2). The formation is 2900 feet thick in the type section and may be as thick in the Coopers Lake quadrangle several miles south of the exposures in the Saypo area. In the latter region, however, the lower part of the formation is cut out by the Lewis overthrust which threw the Miller Peak on Mississippian limestone north of Deer Creek and on shale of the Kootenai formation south of Deer Creek (Fig. 2). The maximum thickness of the Miller Peak argillite exposed in the area is estimated to be 1000 feet. On the ridge north of Deer Creek the lower 300 feet of beds exposed is pale-green and maroon, thin- to fissile-bedded, hard siliceous argillite overlain by 500 feet of alternating red and green, finely arenaceous, ripple-marked argillites; ½- to 1½-inch beds of gray fine-grained quartzite; and a few 1- to 4-inch beds of buff-tan, fine-grained, dense, slightly dolomitic and quartzitic sandstone. The upper 300 feet of the Miller Peak consists of tan-weathering dolomitic and quartzitic sandstone interbedded with red and green, finely arenaceous argillite which together form a transition zone from the dominantly clastic rocks beneath into the limestones and dolomites above. The Miller Peak argillite weathers to pale-maroon steeply rounded slopes and forms the lower end of the ridges north and south of Deer Creek on the west side of the South Fork of Sun River Valley.

CAYUSE LIMESTONE

General statement.—The Cayuse limestone crops out in the southwest part of the Saypo quadrangle. The entire thickness of the formation is exposed throughout a distance of only 4 miles on the upper part of the hills west of the South Fork and south of the West Fork of Sun River (Fig. 2). North and south of Deer Creek the Cayuse limestone rests conformably on the Miller Peak argillite, but north of the West Fork of Sun River the Cayuse is thrust upon Hannan limestone (Mississipian) and shale of the Kootenai (Lower Cretaceous) along the Lewis overthrust. North of Prairie Creek the Cayuse limestone is cut out by the Lewis thrust so that sandstone of the Hoadley formation rests against shale of the Kootenai. The name Cayuse is taken from Cayuse Creek, a tributary of the South Fork of Flathead River in the northern part of the Ovando quadrangle where the formation is thickest (2100 feet). The description and name of the Cayuse limestone has not been published previously, although the formation has been studied and mapped since 1930 in the Coopers Lake, Ovando (Clapp, M. M., 1936, p. 18), and Silvertip quadrangles. The detailed measured type section will be published at a future time.

Original description.—In the Saypo quadrangle on the ridge north of Deer Creek the entire thickness of the Cayuse limestone is well exposed, but precise measurement would be difficult because some of the beds are closely folded and mashed. The formation was not measured, and the thickness given here (1000 feet) was scaled from the map and computed from a reconnaissance traverse across the strike. On the ridge, the lower fourth of the Cayuse consists of dull- to pale-gray, finely crystalline, siliceous, buff-tan-weathering dolomite interbedded with pale-maroon and green-gray fissile argillite, and several 8- to 12-inch beds of gray, vitreous, coarse-grained quartzite. The rocks weather buff in contrast to the underlying maroon Miller Peak argillite. Some of the dolomites in the lower part of this unit contain irregular flakes and stringers of clay which cause them to weather in relief and to simulate "Molar tooth" structure (Daly, 1912, p. 74). Overlying the lower unit of the Cayuse is 100 feet or more of gray dolomite which weathers brown and is interbedded with black-gray fissile argillite.

The third unit of the formation consists of 250 to 300 feet of alternating shaly-bedded, calcareous and dolomitic argillite, and 2- to 12-inch beds of gray, finely sideritic and dolomitic, hard marble marked with fine wavy bands. The upper 400 feet of the Cayuse limestone is composed of blue- and dull-gray, thick-bedded, finely oölitic, coarsely crystalline limestone which contains glassy quartz grains, weathers dull gray, and passes upward into pale-gray thin and irregularly bedded pure fine-grained algal limestone. Locally, the algal "heads" are 6 to 30 inches in diameter and form typical reefs (bioherms). The limestones in this unit weather gray and are the purest in the formation. The contact of the Cayuse and the overlying Hoadley is covered, but the two formations probably intergrade lithologically through interbedded argillite and limestone, the argillite increasing upward in proportion to the limestone. The contact between the formations was arbitrarily drawn at the base of the first prominent argillite.

The Cayuse limestone is so resistant to weathering that it forms steep slopes and rounded knobs above the Miller Peak argillite. The buff and gray color to which the Cayuse weathers forms a conspicuous band between the maroon of the Miller Peak and the deeper red of the Hoadley formation.

HOADLEY FORMATION

Original description.—The Belt rocks above the Cayuse in the Swan Range in the northern part of the Ovando quadrangle 16 to 20 miles southwest of the Saypo quadrangle are approximately 11,000 feet thick if 5000 feet be subtracted for the thickness of the combined Miller Peak and Cayuse (Deiss, 1935, p. 107-108). In the Saypo quadrangle the

combined thickness of the Hoadley and Ahorn formations is estimated to be 6200 feet, of which the Hoadley is 4100 feet. Because the Hoadley is cut by three low-angle thrusts, the exact thickness of the formation is not measurable in the southwest part of the Saypo quadrangle. The following thicknesses of different parts of the formation are estimates scaled from the map and computed from field notes made during reconnaissance traverses across the strike northeast of Hoadley Reef and southwest of Deer Creek (Fig. 2). The formation is named from Hoadley Reef which is less than a mile southwest of the largest exposure of these rocks in the Saypo quadrangle (Pl. 5, fig. 2).

The lower 400 feet of the Hoadley formation is composed of dark-green, green-gray, and some maroon, thick-bedded, siliceous argillite; and argillaceous, vitreous, fine-grained quartzite in beds 1 to 12 inches thick. Many beds are finely banded. The rocks weather drab gray and dull green above the buff-gray weathered Cayuse limestone. Above the basal argillites is 1000 to 1050 feet of buff, arenaceous, calcareous argillite and fine-grained sandstone which grades upward into pale-green-buff-weathering, calcareous sandstone and argillite 500 feet thick. Above this lower buff unit is 1500 feet of pink, red, and occasional buff sandstones and argillites interbedded (Pl. 5, fig. 2). Clay galls and ripple marks are numerous on some beds in this unit. The second or stratigraphically higher buff-weathering unit is approximately 650 feet thick and lies upon the red sandstone and argillite. The lower 300 feet of this buff unit consists of buff-weathering, calcareous, sericitic and limonitic argillite and a few intercalated beds of red, argillaceous sandstone. The upper 350 feet is composed of buff-weathering, calcareous argillite and argillaceous limestone interbedded with pink-gray, ripple-marked quartzite and occasional 10- to 18-inch beds of edgewise conglomerate. In the lower part are 6- to 18-inch beds of Algae (*"Collenia"*).

The upper 700 feet of the Hoadley formation consists of dark- and brilliant-red, soft, thin- and thick-bedded, platy to ripple-marked sandstone interbedded with some pale-red, arenaceous argillite and a few beds of tan and buff sandstone. The rocks in this interval weather deep red and are conspicuous throughout the western part of the quadrangle where the Hoadley usually forms gentler slopes below the Ahorn quartzite.

AHORN QUARTZITE

Original description.—The youngest Belt formation in the Saypo quadrangle is exposed along the extreme western part of the map area and is completely represented only in the southwestern part of the quadrangle where it underlies the basal Cambrian sandstone on the northeastern

slopes of Hoadley Reef (Pl. 5, fig. 2). The formation has not been measured but is estimated to be 2100 feet thick north of Hoadley Reef. Because the upper part of the Ahorn was greatly eroded during the Lipalian interval (Deiss, 1935, p. 111-113) the thickness of the formation varies considerably within relatively short distances. The formation is named from Ahorn Creek in the southeastern corner of the Silvertip quadrangle where the headward part of the creek is incised in these rocks.

The Ahorn in the Saypo quadrangle is composed of two sharply demarcated units, a lower, pink quartzite 1700 feet or more thick, and an upper, green and red argillite 400 feet thick. The lower three-fourths of the basal unit consists of pink and pale-maroon, thick-bedded, medium- to coarse-grained, arenaceous and pure quartzite which in some beds is cross-bedded and spotted lighter pink. The upper fourth of the unit is composed of lavender-pink, coarser-grained, thicker-bedded, and more strongly cross-bedded, vitreous quartzite.

The upper 400-foot unit is composed of interbedded, green, green-gray, and a little red-gray, thin- and fissile-bedded argillite and a few intercalated thin beds of fine-grained sandstone. These rocks weather to steep, smoothly rounded slopes and are directly overlain by pebbly sandstone of the basal Cambrian.

CAMBRIAN SYSTEM

GENERAL STATEMENT

Cambrian rocks are exposed in only three small areas in the southern half of the Saypo quadrangle: on Hoadley Reef in the extreme southwest corner of the quadrangle, between the heads of Bear and Moose creeks, and in a long narrow belt one-fourth to three-fourths of a mile wide extending from Allan Mountain northward at least 15 miles (Fig. 2). In the Lewis and Clark and Sawtooth ranges both the Middle and Upper Cambrian sediments are recorded by their contained fossils. The youngest Cambrian formation in the area, the Devils Glen dolomite and the upper part of the Switchback shale, probably are St. Croixan in age.

The Middle Cambrian rocks of northwestern Montana have been more carefully and extensively studied than those of any other of the Paleozoic systems in the Cordilleran region. Since 1931 the writer measured 10 detailed sections in the Lewis and Clark Range, and 24 other sections at widely spaced points in the Cordilleran trough from Nevada to British Columbia. The fossils collected from and zoned within each of the measured sections have been studied sufficiently to indicate at least

seven faunal zones in the Middle Cambrian rocks in northwestern Montana (Deiss, 1939, p. 53-55). The Cambrian rocks in the Lewis and Clark and Sawtooth ranges are divided into nine formations of which eight have been recognized in the southern half of the Saypo quadrangle. The only place in the map area, however, where the entire thickness is completely exposed is on Hoadley Reef (Pl. 5, fig. 2). The Cambrian formations and their contained trilobites have been adequately described (Deiss, 1939, p. 34-130) and, therefore, only those exposed in the southern half of the Saypo need to be summarized here.

The names of the Cambrian formations used here apply only to northwestern Montana because the rocks are different lithologically, are much thicker, and contain different faunal facies than those in central Montana. The names of the Cambrian formations in the standard sections in central and southern Montana already are revised and established.

FLATHEAD SANDSTONE

The basal Cambrian formation in the area is the Flathead sandstone which is 50 to 100 feet thick. The Flathead consists of white quartz sand cemented in a matrix of siliceous and hematitic clay which causes dull-red stains. The most diagnostic feature of the lower beds, if not of the whole formation, is the presence of white and tan pure quartz pebbles which range from ⅛ to 3 inches in diameter. The coarser-grained rock grades upward into fine-grained sandstone, which contains occasional intercalated thin beds of shale in the upper part of the formation and at the top usually a thin zone of impure hematite. The Flathead sandstone in the southern part of the Saypo quadrangle is exposed only on the lower slopes on the northeast side of Hoadley Reef (Fig. 2) where it rests slightly unconformably on the green argillite of the Ahorn formation and forms ledges or low cliffs below the smoothly weathered slopes of Gordon shale. Limonite grains are abundant in the sandstone and cause it to weather to the buff-tan color characteristic of the Flathead sandstone throughout the region.

GORDON SHALE

The Gordon shale, approximately 220 feet thick, is composed of drab-green-gray and maroon, fissile to chunky, micaceous shale interbedded in the upper-middle part of the formation with thin beds of sandstone and thicker beds of limestone. The shale weathers green, maroon, and gray, and its soft character produces gentle slopes and a bench above the Flathead sandstone. The upper shales of the Gordon contain numerous fossils, the most diagnostic of which are trilobites. The more

common species of trilobites and brachiopods in the formation are (Deiss, 1939, p. 38; Bell, 1941, p. 212, 220, 242):

Acrothele panderi Bell
Alokistocare charax (Walcott)
Anoria baton (Walcott)
Clavaspidella bela (Walcott)
Glossopleura belesis (Walcott)
Iphidella nyssa Walcott
Strotocephalus gordonensis Resser
Wimanella simplex Walcott
W. cf. *W. simplex* Walcott
Zacanthoides cnopus Walcott

DAMNATION LIMESTONE

Resting conformably on the Gordon shale throughout the Lewis and Clark and Sawtooth ranges is approximately 1300 feet of limestone and a few intercalated shales. These calcareous rocks contain five distinct trilobite faunas and have been divided into six formations of which five are exposed in the southern part of the Saypo quadrangle on Hoadley Reef. The oldest of the formations is the Damnation limestone (Deiss, 1939, p. 38-40) which is approximately 150 feet thick in the Saypo map area. The Damnation consists of a basal unit 20 to 30 feet thick of blue- and tan-gray, irregularly and thin-bedded, occasionally oölitic, fossiliferous limestone which contains much limonitic, bright-buff-weathering clay. These rocks break down to smooth, buff talus slopes above the Gordon shale and below the massive upper part of the Damnation limestone. The upper and thicker part of the Damnation is composed of chocolate-gray and tan, fine-grained, thick- and thin-bedded, hard massive limestone, which contains flakes and blebs of buff and orange-colored, siliceous clay. These rocks form sheer cliffs wherever the limestone was sculptured by water or ice. The basal part of the Damnation contains trilobites of the *Glossopleura* fauna, which is known from central Nevada to British Columbia. The species of trilobites and brachiopods characteristic of the Damnation limestone are (Deiss, 1939, p. 39; Bell, 1941, p. 212, 239):

Glossopleura alta Deiss
G. fordensis Deiss
G. inornata Deiss
G. minima Deiss
G. perryi Deiss
G. thomsoni Deiss
Iphidella cf. *I. nyssa* Walcott
Kootenia erromena Deiss
K. exilaxata Deiss
K. fragilis Deiss
K. infera Deiss
K. scapegoatensis Deiss
Nisusia 2 new species

DEARBORN LIMESTONE

The Dearborn limestone consists of a lower, shaly unit, and an upper and much thicker limestone unit. The basal part of the formation is composed of green, fissile shale, and a few intercalated 1- to 3-inch beds of limestone conglomerate overlain by alternating green, fissile shales and thin limestones. Well-preserved trilobites are numerous 5 to 15 feet above the base. The lower unit is approximately 35 feet thick. The upper part of the Dearborn is composed of chocolate- and tan-gray, fine- to medium-grained, thick- and thin-bedded limestone,

which contains varying amounts of buff and orange-tan clay in the form of flakes, nodules, and partings. The lower unit of the Dearborn forms a bench or slope between the sheer cliffs of the upper part of the Dearborn and those of the upper Damnation on the north side of Hoadley Reef. The Dearborn has not been seen elsewhere in the southern part of the Saypo quadrangle. The formation is approximately 300 feet thick. The trilobites found in the lower part of the Dearborn have not been studied carefully but clearly represent a short-lived, recurrent *Glossopleura* assemblage (Deiss, 1939, p. 52, 54-55) similar to *Glossopleura belesis* (Walcott) in the upper part of the Gordon shale.

PAGODA LIMESTONE

Throughout the Lewis and Clark and Sawtooth ranges the Dearborn limestone is overlain by limestone and shale which averages 300 feet thick and is named the Pagoda limestone (Deiss, 1939, p. 40-41). The lower fourth or fifth of the Pagoda consists of a basal, thin-bedded, fossiliferous limestone interbedded with green, fissile shales and several 2- to 4-inch beds of intraformational conglomerate. The upper four-fifths of the Pagoda is composed of cream- and chocolate-gray, finely crystalline limestone, thin-bedded in the lower and upper parts, and oölitic and thick-bedded in the middle part. The upper limestone of the Pagoda forms the boldest and sheerest cliffs of any Cambrian rocks in the area. The only places the Pagoda was seen in the southern part of the Saypo quadrangle are on the north side of Hoadley Reef (Pl. 5, fig. 2) and on the west side of Big George Gulch where the rocks are much folded and sheared. At the latter locality, only the upper part of the Pagoda is exposed and contains more green shale than in the region to the south and west. The possibility exists that the Pentagon shale (Deiss, 1939, p. 42-44) is present in the Big George exposure and that the beds tentatively assigned to the Pagoda may belong to the Pentagon. The lower beds of the Pagoda contain numerous trilobites of the *Glyphaspis* fauna which seems to be more prolific in some places than in others. The common species of trilobites and brachiopods in the fauna are (Deiss, 1939, p. 41; Bell, 1941, p. 210, 214):

Bolaspis globulifera Deiss
B. grandis Deiss
B. minuta Deiss
B. vera Deiss
Glyphaspis delicata Deiss
Glyphaspis paucisulcata Deiss
G. robusta Deiss
G. storeyi Deiss
Icadonta? cf. *I. nizonensis* Bell
Iphidella hexagona Bell

PENTAGON SHALE

In the north-central part of the Lewis and Clark Range, the Pagoda limestone is conformably overlain by as much as 290 feet of buff-gray, platy, calcareous, thick-bedded, extremely fossiliferous shale, which

grades upward into thin- and thick-bedded limestones. The shale (Diess, 1939, p. 42-44) thickens northward, indicating clearly that the muds were deposited along the southern shore of Montania from which they were eroded (Deiss, 1941, p. 1095-1097). The Pentagon shale contains the most prolific fauna of any Middle Cambrian formation in Montana. The formation is not present on Hoadley Reef nor along the west edge of the map area at the head of Bear and Glenn creeks. The Pentagon also probably is absent on the west side of Big George Gulch in the central part of the map area. If these beds, tentatively assigned to the upper part of the Pagoda, are Pentagon the formation has changed lithologically between the head of Lick Creek in the western part of the Silvertip quadrangle and the central part of the Saypo quadrangle.

STEAMBOAT LIMESTONE

The Steamboat limestone rests upon the Pagoda in the southern part of the map area where it is well exposed on Hoadley Reef. The Steamboat is also exposed on the east side of White Ridge (Fig. 2), along the west side of Big George Gulch, at the head of Cabin Creek, and on the west side of Leavitt Creek Valley. The formation (Deiss, 1939, p. 45) averages approximately 275 feet in thickness in the measured sections in the Lewis and Clark Range and is 239 feet thick at its type locality on Prairie Reef, a mile west of the Saypo quadrangle. The Steamboat consists of pale-gray and tan, fine-grained, thick-bedded limestone in the lower part, overlain by interbedded tan-gray, irregularly bedded, fossiliferous limestone and green fissile shale in alternating units. The upper 75 feet of the formation in the type locality is composed of gray and tan, crystalline, thick- and thin-bedded limestone, which contains flakes and nodules of tan clay. The Steamboat limestone forms bold cliffs on Hoadley Reef (Pl. 5, fig. 2) and on the west side of White Ridge at the head of Bear and Glenn creeks, and also forms the upper part of the 8100-foot peak at the head of Circle Creek (Fig. 2). The green, intensely folded shales on the west side of Big George Gulch in Elbow Gorge are part of the Steamboat limestone unit. The thin-bedded, lenticular limestones, interbedded with the green shale 80 feet above the base of the formation, contain many individuals of the characteristic *Thomsonaspis* fauna (Deiss, 1940, p. 783, 785) of which the most common species of trilobites are (Deiss, 1939, p. 45):

Coelaspis prima Deiss
Glossocoryphus cliffensis Deiss
G. typus Deiss
Glyphaspis dearbornensis Deiss
G. indenta Deiss
G. similis Deiss
Kochaspis dearbornensis Deiss
K. resseri Deiss
K. unzia (Walcott)
K. upis (Walcott)
Mcnairia inornata Deiss
Thomsonaspis obscura Deiss
T. striata Deiss

SWITCHBACK SHALE

The Steamboat limestone is overlain conformably by 75 to 125 feet of green and gray, soft, slightly arenaceous and locally calcareous, fissile to chunky shale, interbedded with gray, crystalline, flaggy-bedded magnesian limestone which weathers rusty tan. In the upper part of the formation are a few thin beds of limestone-conglomerate. The Switchback shale weathers to green-buff or bright-brown slopes in some places but is often covered with talus of Devils Glen dolomite. On Hoadley Reef and at the head of Cabin Creek in the Saypo quadrangle the Switchback forms a bench and gentle slope below the sheer light-gray cliffs of dolomite (Pl. 5, fig. 2). Norman Denson is studying the trilobites of the Switchback shale. Concerning the age of the fossils Denson (personal communication) says:

"In Dearborn Canyon (Coopers Lake quadrangle) trilobites were collected in talus derived from and resting upon the Switchback shale. These collections contain elements of both the youngest Middle Cambrian fauna (*Deissella-Centropleura* fauna) and the oldest Upper Cambrian fauna (*Cedaria* fauna). The Middle-Upper Cambrian boundary, therefore, lies somewhere within the Switchback shale. Its exact position within that shale can not be determined in Dearborn Canyon because of poor exposures."

DEVILS GLEN DOLOMITE

General considerations.—The Devils Glen dolomite (Deiss, 1939, p. 46-47), the youngest Cambrian formation in the southern part of the Saypo quadrangle, is composed of white-gray, finely crystalline, massive, thick-bedded dolomite. The Devils Glen is much thicker west and south of the Saypo quadrangle (averages 353 feet) but is prominent on Hoadley Reef, Allan Mountain, and in the upper part of Circle and Cabin creeks, and is well exposed on the west side of Big George Gulch (Fig. 2).

Cambrian section on Allan Mountain.—The lithologic characteristics of the Devils Glen in the south-central part of the Sawtooth Range are given in the following section which was measured in 1940 on the east spur of Allan Mountain in the SW. ¼, sec. 20, T. 21 N., R. 9 W. The base of the section is approximately 7600 feet in altitude and is nearly half a mile east of the Bench Mark (8100 feet) on top of the mountain (Fig. 2).

DEVONIAN. *See* Devonian section on Allan Mountain, in this report.

UPPER CAMBRIAN
Devils Glen dolomite

	Feet	Meters
7. Dolomite: buff-gray mottled salmon-pink, thinner-bedded than below in section and coarser-grained. Weathers white gray and white buff gray, mottled pink on some beds, and to vesicular surfaces. Pink color may be result of small amounts of manganese	37	11.3

	Feet	Meters
6. Dolomite: pale- to darker-tan, mottled dull-pink in small irregular areas. Weathers pale buff gray and to irregular surfaces. Upper bed 3½ feet thick	5	1.5
5. Dolomite: white-gray, slightly vitreous, finer crystalline than in underlying unit No. 4, in beds as much as 3 feet thick. Upper beds mottled pink tan and salmon pink in areas as much as 6 inches in diameter	59	17.9
4. Dolomite: pale-gray, finely crystalline, sugary. Contains small amount of buff clay as flakes and blebs in lower 10 feet. Upper beds purer dolomite, thicker-bedded, and weather to faint-buff stains. All weather white gray and to sugary surfaces and to large subangular boulders. Joints weather to slight depressions. Forms cliffs or steep slopes wherever cut by ice	97	29.9
Total thickness of Devils Glen dolomite	198	60.6
Switchback shale		
3. Shale: almost identical to that below in No. 1	2-3	0.9
2. Dolomite: bright-orange-buff, in irregular beds as much as 1½ inches thick. Contains flakes and partings of buff clay, and white calcite geodes and veinlets 1 to 2 mm. thick. Weathers bright buff	2.5	0.7
1. Shale: pale-green and pale-green-gray, fissile, soft, extremely fine-grained. Weathers dull green	10	3.0
Total thickness exposed of Switchback shale	15.5	4.6
Total thickness of Cambrian exposed in measured section	213.5	65.2

At this point a high-angle thrust fault cuts off the section and throws Switchback shale against Lone Butte limestone member (Devonian).

DEVONIAN SYSTEM

GENERAL STATEMENT

The Devonian rocks in the Saypo quadrangle are approximately 1000 feet thick, are all Upper Devonian [Naples? (G. A. Cooper, personal communication) and Chemung], and cannot be correlated precisely upon evidence from fossils with the Jefferson and Three Forks formations of Peale (1893, p. 27-32) in central Montana. These rocks in the Sawtooth Range also are not lithologically similar to the Three Forks shale or to the typical Jefferson limestone.

In 1932 the writer measured seven stratigraphic sections of the Devonian sediments in the Swan and Lewis and Clark ranges and divided the rocks, then thought to be equivalent to the Jefferson, into five lithologic members (Deiss, 1933, p. 41-44), which, from the base upward, were designated the: White Ridge limestone, Glenn Creek shale, Coopers Lake limestone, Lone Butte limestone, and Spotted Bear limestone. From the measured

sections south and west of the Saypo quadrangle, the following species, identified by G. A. Cooper (personal communication), were found:

Alveolites sp.
Ambocoelia sp.
Athyris sp.
Atrypa hystrix var. *spinosa*
A. "missouriensis"
A. sp.
Cladopora sp.
Cyrtospirifer sp.
Favosites sp.
Pachyphyllum sp.
Productella sp.
Schizophoria sp.
Schuchertella sp.
Spirifer argentarius
S. engelmanni?
S. mesacostalis type
Stromatopora sp.
Stropheodonta sp.

Concerning the age of the Devonian rocks in northwestern Montana and the fossils collected from them Cooper (personal communication) says:

"Much of the collection is too poor for identification and most of the species are new. It is my belief that the sequence is not typical Jefferson and it is not Three Forks as hitherto reported. The Survey regards the Jefferson as of Middle Devonian age. This in my opinion is an error; my findings indicate that typical Jefferson correlates with the lower Devils Gate of Merriam and is probably of Naples age. The Coopers Lake member fauna contains a *Spirifer* related to *S. mesacostalis* and an *Atrypa* related to *A. hystrix*, together with *Pachyphyllum*, which to my knowledge does not go above the Chemung. The Jefferson also contains *Pachyphyllum*. It appears to me that what is called Three Forks limestone (as opposed to shale) is probably of Chemung age and also pre-Three Forks."

In the Sawtooth Range in the southern part of the Saypo quadrangle only three of the five members of Devonian rocks are represented (White Ridge, Coopers Lake, and Lone Butte), and they are abnormal in composition and thickness. The upper part of the rocks correlated with the Lone Butte member contains much more argillaceous material than in the Lewis and Clark Range.

All the Devonian rocks in the southern part of the Saypo quadrangle, except an area of less than 1 square mile near the head of Glenn Creek, are confined to the Sawtooth Range. The rocks crop out in three belts: (1) along the north front of Sawtooth Ridge, the east side of Home Gulch, and the east face of Castle Reef (Fig. 2); (2) a much narrower belt along the west side of Mortimer Gulch and directly south on the west side of the gulch west of Beaver Creek Valley; and (3) a well-exposed belt which extends from the south edge of the quadrangle north along the east side of Allan Mountain, the west side of Leavitt Creek and Big George Gulch, across the heads of Circle and Cabin creeks, and continues an unknown distance northward.

The Devonian rocks in the Saypo quadrangle probably must be given a new name or names, but the mapping is not complete enough to determine whether they comprise one or two formations. For this reason assignment of new names to the tentative Devonian formations is deferred until the rocks in the quadrangle are completely mapped.

DEVONIAN SECTION ON ALLAN MOUNTAIN

The lithologic characteristics of the Devonian rocks in the southern part of the Sawtooth Range are given in the following section which was measured on the same east spur of Allan Mountain as was the preceding section of the Devils Glen dolomite. Most of the section is in the SW. ¼, sec. 20, T. 21 N., R. 9 W., but the upper part lies in the eastern side of the SE. ¼, sec. 19, T. 21 N., R. 9 W.

	Feet	Meters
MISSISSIPPIAN (Hannan limestone)		
Intraformational breccia in lower 3 to 7 feet overlain by gray, crystalline, hard, thick-bedded limestone which weathers dull gray. Forms low cliff in slope here and is base of Hannan limestone.		
DEVONIAN		
15. Limestone and dolomite: pale-tan and buff-gray, finely to medium-crystalline, hard, dense dolomite in 1- to 2-foot beds, and more bright-buff-weathering, lumpy, argillaceous limestone and dolomite. Rocks weather bright-buff	106	32.3
14. Limestone: pale-gray, finely crystalline, thick- and thin-bedded, dolomitic. Some beds finely banded, others platy, finely arenaceous, and very slightly vitreous	51	15.5
13. Limestone and dolomite: tan-gray, thick-bedded, vitreous, petroliferous, finely arenaceous, dolomitic limestone in 4- to 7-inch beds; contains many blebs and flakes of limonitic orange-tan clay. Thicker units of drab-tan rubbly and shaly thin-bedded buff-tan-weathering limestone, and intercalated 1- to 4-foot units of pale-gray limestone which forms ledges here, but is covered with talus of softer, drab-tan beds on surrounding slopes	118	39.9
12. Limestone: alternating units of light-gray, vitreous, slightly siliceous limestone; buff, shaly limestone; and thick beds of drab-tan, rubbly argillaceous, fetid limestone. Weathers to alternate gray, buff, and tan bands on hillside	104	31.7
11. Dolomite and limestone: drab-brown and pale-tan, finely to medium-crystalline, thick- and thin-bedded, very slightly vitreous dolomite. Some beds rubbly and appear as rolled mud; others finely banded. Light-gray-weathering limestone in upper part	204	62.2
10. Limestone: light-gray, finely crystalline, thin-bedded. In basal 3 to 5 feet, one reef or bed contains angular fragments of drab-tan limestone ¼ to 4 inches in diameter, and numerous poorly preserved *Stromatopora*. Most beds thin, and all weather white gray slightly banded	31	9.4
9. Dolomite: drab-tan and brown, vesicular, fetid, finely arenaceous, in beds as much as 4½ feet thick; weathers drab brown and to rounded sugary surfaces	82	25.0
8 Limestone: dull-gray, medium-crystalline, hard, in ⅛- to 4-inch beds. Thick beds contain dark- and light-gray chert lenses 4 to 7 inches long and ½ to 1½ inches thick. Limestone weathers pale-gray in contrast with drab-brown beds above and below	5	1.6
7. Limestone: drab-tan, slightly vitreous on fresh fracture, slightly vesicular, and probably finely arenaceous. All beds more or less dolomitic. Weathers dull tan and to rounded sugary surfaces	27	8.2

	Feet	Meters
6. Dolomite: buff-gray, argillaceous, finely arenaceous, thin-bedded, platy; weathers to gray and buff-gray angular fragments	2	0.6
Thickness * of units 6 to 15	730	226.4
* Note: Units 6 to 15 are equivalent to the Lone Butte limestone member in the Lewis and Clark Range (Deiss, 1933, p. 43-44).		
5. Limestone: pale- and dark-tan, finely to medium-crystalline, hard, thick-bedded, massive. Many thick beds in upper two-thirds mottled light- and dark-gray and contain *Stromatopora* 2 to 7 inches in diameter. Lower 8 to 12 feet thin-bedded, and contains *Atrypa* cf. *A. hystrix, Favosites* sp. and other Devonian fossils	71	21.6
4. Limestone: pale-tan-gray, and occasional dark-tan bed, finely crystalline, thin- and thick-bedded, hard. Limestone weathers pale gray and forms cliffs. Many *Stromatopora* ½ to 1½ inches in diameter, and fragments of *Atrypa?* in upper and middle parts	97	30.5
Thickness * of units 4 and 5	168	52.1
* Note: Units 4 and 5 are equivalent to the Coopers Lake limestone member in the Lewis and Clark Range (Deiss, 1933, p. 43).		
3. Limestone, dolomite, and shale: dull-buff-tan, argillaceous, dolomitic, shaly limestone and dolomitic shale. Passes upward into drab-tan, fetid, finely arenaceous, argillaceous dolomite in 1- to 4-inch beds which contain 1- to 8-mm. spots of green dolomite or clay. Upper fourth of unit covered with float and soil	55	16.8
2. Dolomite: buff, argillaceous, 1- to 5-inch beds; weathers buff and forms ledge here	2	0.5
1. Dolomite, limestone, and shale: bright-straw-buff-weathering, argillaceous, dolomite and dolomitic shaly limestone in ⅛- to 1-inch beds, and partings of buff shale. Forms saddle on ridge here and buff band above white-weathering dolomite on mountain side	25	7.6
Thickness * of units 1 to 3	82	24.9
* Note: In this section the maroon beds of the Glenn Creek shale member are absent, but lower on this ridge in the underlying fault wedge a few beds are mottled purple and maroon. Units 1 to 3 may be equivalent to the White Ridge limestone member in the Lewis and Clark Range (Deiss, 1933, p. 42).		
Total thickness of Devonian	980	303.4

CAMBRIAN *(Devils Glen dolomite)*. *See* Cambrian section on Allan Mountain.

Fossils are not readily obtainable from the Allan Mountain section, but on the north end of Sawtooth Ridge 5½ miles east of Allan Mountain the following species, identified by Cooper and Kirk (personal communication), were collected: *Ambocoelia* sp., *Spirifer* of *S. mesacostalis* stock, *Cyrtospirifer* of *C. whitneyi* stock, and *Chonetes* sp. or *Schuchertella* sp.

CARBONIFEROUS SYSTEM

GENERAL STATEMENT

The most prominent cliff-forming rocks in the southern part of the Saypo quadrangle in the Sawtooth Range are Mississippian limestones which weather white gray and form the crest of nearly all the high ridges

and most of the highest peaks in the range. Most of the Mississippian limestone in the quadrangle probably is equivalent to the Madison limestone of Peale (1893, p. 33-39), but the upper beds contain Meramec and Chester species and must be upper Mississippian. Stebinger (1918, p. 155) first noted the presence of upper Mississippian limestone in the Saypo quadrangle, and Girty (Stebinger, 1918, p. 155) correlated it with the Brazer (Richardson, 1913, p. 313-314) limestone of northeastern Utah. Fossils have not been collected from a carefully measured section in the Sawtooth Range, and the writer was unable to separate the lower and upper Mississippian rocks into two mappable units in the southwestern part of the Saypo quadrangle. As a result, the boundary between the Madison and younger Mississippian rocks is unknown. When the boundary is known the Mississippian limestones in the southern part of the quadrangle still must be mapped as one formation.

In 1932 Deiss (1933, p. 14, 16, 21-22, 29) measured four sections, and in 1933 measured two additional sections (unpublished) of Mississippian rocks in the Flathead and Lewis and Clark ranges and collected fossils from them. In the sections all rocks were assigned to the Madison limestone which was divided into five members. From oldest to youngest these members are (Deiss, 1933, p. 45-48): (1) Silvertip conglomerate, (2) Saypo limestone, (3) Dean Lake chert, (4) Rooney chert, and (5) Monitor Mountain limestone. Rocks equivalent to the basal Silvertip conglomerate are present in the Saypo area on Allan Mountain (Pl. 1, fig. 2) and on the ridge west of Big George Gulch. The conglomerate is readily separable from the overlying Saypo limestone, which is also represented in the Sawtooth Range. The Saypo limestone, however, is not clearly demarcated from the overlying Mississippian limestone. The Dean Lake and Rooney chert members are not typically developed in the southern half of the Saypo quadrangle where the black-gray chert and limestone characteristic of the Dean Lake chert member are absent. The 355 feet of limestone immediately above the beds correlated with the Saypo limestone member may be equivalent to the Dean Lake chert member. However, the limestone is much darker and the chert is smaller in amount and much paler gray than in the Dean Lake chert member to the west and south. The Monitor Mountain limestone has never been found except in the type locality in the Coopers Lake quadrangle.

HANNAN LIMESTONE

General considerations.—The Mississippian rocks in the Saypo quadrangle must be given a new name because the rocks are not precisely equivalent in age to the Madison or to the Brazer limestones but are

partly correlative with both. The name Hannan limestone is proposed for all the strata which lie between the Devonian and Jurassic rocks in the map area. The name is taken from Hannan Gulch (Fig. 2) where the Mississippian rocks form the crest of both sides of the gulch (Fig. 3). Almost the complete section of the Hannan is exposed on the north side of Sun River Canyon at the south end of Castle Reef which forms the east side of Hannan Gulch.

A complete stratigraphic section of the Hannan limestone has not been measured in the Sawtooth Range, but the general lithologic characteristics of the formation can be given from data acquired while mapping the rocks during 1940.

Original description.—The lower 30 feet of the Hannan limestone consists of 3 to 7 feet of intraformational breccia (Pl. 1, fig. 2) composed of dull-tan limestone fragments 1/16 to 2½ inches in long diameter embedded in a matrix of tan and dull-gray limestone, overlain by 23 to 27 feet of gray, medium-crystalline, thick-bedded hard limestone, which weathers pale gray. The unit is correlated with the Silvertip conglomerate member of the Madison limestone in the Lewis and Clark Range.

Above the basal unit is approximately 120 feet of dull-gray, shaly-bedded, finely crystalline, argillaceous limestone, which contains partings and flakes of buff clay. The limestone weathers to angular bright-buff fragments. The upper part of the unit is alternately thick- and thin-bedded. This buff-weathering shaly unit is similar in composition and occupies the same stratigraphic position as the Saypo limestone member in the Lewis and Clark Range.

Dark-tan, finely crystalline, generally thin-bedded, and some thick-bedded limestone, 355 feet thick, overlies the buff-weathering limestone and contains irregular nodules and small lenses of pale-gray chert. Crinoidal fragments and imperfectly preserved brachiopods and corals are sparingly distributed throughout the upper half of the unit. The limestones weather drab and dull gray.

The 665 feet of the Hannan limestone above the dark-tan limestone forms a readily recognizable unit in the southern part of the Saypo quadrangle. The lower 450 feet of the unit consists of pale-tan and dark- and light-gray, coarse-grained, thick-bedded limestone, which contains white and pale-gray chert in 2- to 3-inch irregular beds and irregular nodules of all sizes ranging from less than 1 to as much as 11 inches in diameter. Crinoidal fragments are fairly common in many beds, and in the lower part of the unit are concentrated into ½- to 2½-inch lenses composed largely of crinoid columnals. The limestones in this unit contain the

most abundant fossils in the Hannan limestone. In the lower 2 feet individuals of a small species of *Zaphrentis* and a large *Cyathophyllum?* are very abundant, and fewer but numerous individuals of *Syringopora* also occur. Numerous shells of *Spirifer* and other brachiopods are associated with corals, and together these fossils mark the zone which is recognizable at many places in the Sawtooth Range. The upper 215 feet of the 665-foot unit is composed of pale-gray, coarse-grained, thin- and some thick-bedded, massive, cliff-forming limestone which weathers white gray and contains irregular siliceous lenses or nodules. Pure chert is rare in these beds, which may comprise the upper part of the Madison limestone in the southern part of the Sawtooth Range. The unit probably is equivalent to most of the Rooney member in the Lewis and Clark Range.

In 1933 and 1934 Blackstone (1934, p. 85-87) identified the following species of brachiopods which the author collected from the Rooney and Dean Lake chert members in the Flathead and Lewis and Clark ranges south and west of the Saypo quadrangle (Fig. 1). Most of the species listed probably are present also in the lower 1000 feet of the Hannan limestone.

Ambocoelia minuta White
Athyris hannibalensis (Swallow)
A. lamellosa (Leveille)
Brachythyris suborbicularis (Hall)
Camarophorella missouriensis (Winchell)
Camarotoechia choteauensis Weller
C. subglobosa Weller
Chonetes logani Norwood and Pratten
C. ornatus Shumard
C. shumardanus DeKoninck
Cliothyridina glenparkensis Weller
C. incrassata (Hall)
C. obmaxima (McChesney)
C. tenulineata (Rowley)
Delthyris missouriensis Weller
Dielasma formosum (Hall)
Echinoconchus genevievensis Weller
Leptaena analoga (Phillips)
L. convexa Weller
Rhipidomella diminutiva Rowley
R. jerseyensis Weller
R. missouriensis (Swallow)
R. tenuicostata Weller
Spirifer biplicatus Hall
S. biplicoides Weller
S. calvini Weller n. var.
S. forbesi Norwood and Pratten
S. grimesi Hall
S. louisianensis Rowley
S. missouriensis Swallow
S. montgomeryensis Weller
S. mundulus Rowley
S. platynotus Weller
S. striatiformis Meek
S. subaequalis Hall
Spiriferella latior Weller
S. plena (Hall)

The upper 200 feet of the Hannan is dolomite of which the lower 195 feet is white gray, finely crystalline, sugary, thin- and thick-bedded, and contains intercalated, irregular lenses as much as 9 inches thick of drab-gray, slightly coarser-grained, siliceous, tan-weathering dolomite (Pl. 1, fig. 1). The upper 3 to 5 feet of the Hannan is composed of pale-gray, finely crystalline, thick-bedded, extremely vesicular dolomite, which weathers white-gray and to sharp points and cavities on joint and bedding surfaces. Throughout the southern half of the Saypo quad-

FIGURE 1. LENSES OF DRAB DOLOMITE IN UPPER PART OF HANNAN
Length of outlined scale 6 inches

FIGURE 2. ANGULAR LIMESTONE FRAGMENTS IN BASAL HANNAN
Lead pencil (upper center) gives the scale

FIGURE 1. CIRQUES CUT IN DIP-SLOPE OF HANNAN LIMESTONE ON EAST SIDE OF CUT REEF VALLEY
Sawtooth Ridge in background.

FIGURE 2. PERCHED DIORITE ERRATIC ON CREST OF SLICE OF HANNAN LIMESTONE
East side of Arsenic Creek Valley, looking northwest.

CIRQUES AND PERCHED ERRATIC

rangle these dolomites are overlain disconformably by the Ellis formation, but the erosion surface is rarely exposed or conspicuous.

JURASSIC SYSTEM

GENERAL STATEMENT

The Jurassic rocks in the southwest part of the Saypo quadrangle, tentatively assigned to the Ellis formation, are exposed only in the Sawtooth Range where they rest almost conformably upon the slightly eroded upper beds of the Hannan limestone. The Jurassic rocks east of the mountains in the Saypo quadrangle probably are the same as those in the Sawtooth Range. Stebinger (1918, p. 155) assigned all the Jurassic rocks in the quadrangle east of the mountains to the Ellis formation (Peale, 1893, Pl. 1) and described the Ellis in the Saypo area as follows:

> "Immediately overlying the Mississippian limestones in this area, there is a group of black to gray calcareous shales with a few thin, irregular beds of limestone and sandstone, all of which are Upper Jurassic in age and constitute the Ellis formation. The thickness of these shales ranges from 240 to 310 feet and is greatest in the northern part of the area. They form a distinct unit of predominantly shaly beds that is readily separated from the formations above and below it."

AGE OF ELLIS FORMATION IN SAYPO QUADRANGLE

Crickmay (1936, p. 552-555) discussed the problem of the relative ages of the Ellis and Sundance formations in Wyoming and southern Montana and stated that "the faunas of the Ellis are decidedly distinct from, and much earlier than those of the Sundance." Crickmay assigned the Ellis (in terms of European equivalents) to the Callovian, presumably of Buckman, instead of to the Divesian where Reeside (1919, p. 11) placed it. In his summary of the Jurassic of the Yellowstone region Crickmay (1936, p. 550-551) restricted *Gryphaea nebrascensis* Whitfield and Hovey, *Pleuromya weberensis* Meek, and *Pachyteuthis "densus"* (Meek and Hayden) to the Sundance formation and listed *Camptonectes bellistriatus* Meek and Hayden from the upper part of the Sundance in eastern Wyoming. These species are fairly abundant in the Jurassic rocks in the Sawtooth Range. (*See* faunal lists below.)

Concerning the species in the following list, collected in 1940 in the Saypo area, Reeside (personal communication) who identified them says: "This fauna seems to me not older than upper Ellis (Divesian), though some of the species, according to the present records range above and others below the upper Ellis."

The following species were collected at an altitude of 7500 feet on the ridge east of the head of the East Fork of Lange Creek in the SW. ¼, sec. 32, T. 21 N., R. 9 W. The fossils were not zoned in a section

because none was measured there, but all of them probably are limited to the middle 50 to 90 feet of the Ellis formation at that locality.

Astarte n. sp.
A. packardi Meek
Camptonectes cf. *C. bellistriatus* Meek
Cardinia praecisa White
Cucullaea haguei Meek
Dosinia jurassica (Whitfield)
Eumicrotis curta (Hall)
Grammatodon? n. sp.
Gryphaea cf. *G. impressimarginata* McLearn
G. nebrascensis Meek and Hayden
Inoceramus? sp.
Pachyteuthis? sp.
Pinna sp.
Pleuromya weberensis Meek
Quenstedticeras (Eboraciceras?) aff. *Q. hoveyi* Reeside
Q. (E.?) aff. *Q. tumidum* Reeside
Serpula sp.
Trigonoecia sp.
Unicardium? n. sp.
Volsella sp.

In commenting on the fossils collected from locality 2, zone 3, in the Lime Gulch section Reeside says: "Most of the species in this collection have a long range, but the fauna impresses me as belonging high in the interior Jurassic—it could be as high as part of the *Cardioceras* zone (Argovian of European usage)." Concerning the species from the lower part of the same section in locality 2, zone 1, Reeside also says: "This fauna appears to me to be younger than those of the lower Ellis of central Montana. I would place it as not older than upper Ellis (Divesian)."

The evidence from the fossils collected in 1940 seems to indicate that the Jurassic rocks in the southwest part of the Saypo quadrangle are: (1) younger than the Ellis formation in the plains area to the east and southeast, (2) they may be more nearly equivalent in age to the Sundance formation, and (3) they seem to be both Divesian and Argovian in age and may be largely in the *Cardioceras* zone.

Unfortunately, sufficiently detailed stratigraphic and paleontologic work has not been done to determine the vertical range of the species in question and, therefore, to give a final answer to the question of the precise age of the Ellis and Sundance formations. Consequently, the Jurassic rocks in the Saypo area cannot be precisely correlated with either formation. For this reason the name Ellis tentatively is retained for the Upper Jurassic rocks in the southwestern part of the quadrangle.

In summary of the problem Reeside (personal communication) suggests that these rocks represent Jurassic horizons equivalent to the soft, upper, more shaly beds of the Ellis in the Little Rocky Mountains of central Montana, and says concerning the faunas:

"It should be borne in mind that there has been in recent years considerable difference in opinion about ranges of certain common American Jurassic species, and, in my opinion, no one has yet done broad enough work to permit assured decisions about them. Identifications are likely to differ and consequent correlations are likely to differ between students of the Jurassic, and some degree of reservation is still in order in spite of some rather confident published statements."

JURASSIC SECTION AT HEAD OF LIME GULCH

The section was measured in 1940 on the south side of the divide between the heads of Lime Gulch and Cutrock Creek (Fig. 2), at the head of Lime Creek, in the S.W. ¼, sec. 1, T. 20 N., R. 9 W., near the south edge of the Saypo quadrangle at an altitude of approximately 6700 feet. The section was measured east to west and upward. Because the lower part of Ellis and the upper part of the Hannan are covered with soil and float the entire Ellis formation was not measurable.

	Feet	Meters
CRETACEOUS		
Kootenai formation		
23. Sandstone: thin- and thick-bedded (½ to 8 inches), cross-bedded, faint-green-gray; some beds mottled salmon pink; others finely banded with microscopic grains of hematite or magnetite. Weathers yellow green and to rounded surfaces	17	5.2
22. Interval covered. Probably underlain by thin-bedded sandstone and arenaceous shale	54	16.5
21. Sandstone: rusty-brown-weathering, thin-bedded, calcareous	64	19.5
20. Sandstone: tan, thin- and flaggy-bedded, medium- to fine-grained, cement of almost pure calcite. Weathers deep brown and forms ledge here. May form base of Kootenai	5	1.5
Total measured thickness of Kootenai	140	42.7
JURASSIC		
Ellis formation		
19. Sandstone and limestone: olive- and tan-gray sandstone in 1- to 4½-inch beds, contains variable but large numbers of limonite grains. Weathers rusty brown. Occasional bed of manganiferous, argillaceous limestone. This may be basal unit of Kootenai formation instead of No. 20	28	8.5
18. Shale and limestone: tan-gray shale and shaly limestone. Unit partly covered with fragments of calcareous sandstone like No. 19	49	14.9
17. Shale and limestone: olive-tan and gray, calcareous, chunky to fissile shale, and intercalated 2- to 3-inch beds of tan-gray and gray, arenaceous even-bedded, argillaceous, fossiliferous limestone which contains some limonite which weathers rusty on joint faces	35	10.7
16. Shale: olive-green-tan, calcareous, chunky and fissile, very finely arenaceous. Upper 2 feet fissile. Weathers bright buff	9	2.7
15. Shale and limestone: alternating units of gray, fissile, slightly fossiliferous shale and 1- to 1½-foot units of bright-straw-buff-weathering, olive-gray, fossiliferous limestone which contains strongly ribbed pelecypods. Fossil locality 2, zone 4	16	4.9

Fossils:

Astarte packardi White
Camptonectes sp., very small individuals
Grammatodon inornatus Meek and Hayden
Lingula sp.
Nucula sp., probably new
Nuculana sp., probably new
Ostrea strigilecula White
Pleuromya weberensis (Meek)
Thracia aff. *T. weedi* Stanton

Note: Reeside identified the species in this zone and says they could be as young as the *Cardioceras* zone (Argovian) of the Upper Jurassic.

	Feet	Meters
14. Limestone and shale: three 8- to 15-inch units of nodular, gray and tan, extremely argillaceous limestone, alternating with three 2- to 3-foot units of dull- and light-gray, extremely fissile shale. Upper two limestones contain brachiopods, pelecypods, and fragments of ammonites. Fossil locality 2, zone 3. Limestone nodules are gray and lithologic	15	4.6
Fossils:		
Astarte packardi White		
Bittium? sp.		
Camptonectes cf. *C. extenuatus* (Meek and Hayden)		
Corbula n. sp.		
Dentalium? subquadratum Meek and Hayden		
Foraminifer—possibly a nodosarian		
Gryphaea nebrascensis Meek and Hayden		
Inoceramus? sp., a fragment		
Lima cf. *L. cinnabarensis* Stanton		
Pleuromya weberensis (Meek)		
Protocardia cf. *P. shumardi* (Meek and Hayden)		
"Rhynchonella" sp., probably unnamed		
Note: Reeside who identified the species from this zone comments as follows: "Most of the species in this collection have a long range, but the fauna impresses me as belonging high in the interior Jurassic—it could be as high as part of *Cardioceras* zone (Argovian)."		
13. Shale: olive-tan, chunky, thick-bedded, finely arenaceous, slightly calcareous. Weathers tan buff and contains ribbed and smooth pelecypods. Top of locality 2, zone 2	4	1.2
12. Shale: gray, fissile, very slightly calcareous, clayey, soft, smooth, and unfossiliferous	1½	0.5
11. Shale: olive-tan, finely arenaceous, and very calcareous. Contains much finely disseminated limonite which causes rock to weather pale tan; contains fragments of fossils	7	2.1
10. Shale: like No. 9 but more fissile	24	7.3
9. Shale: dull-gray, finely arenaceous, strongly calcareous. Weathers to tiny angular fragments and forms steep slope here	5½	1.7
8. Sandstone or shale: pale-green-tan-gray, argillaceous sandstone or finely arenaceous calcareous shale. Weathers pale tan on joint faces and forms ledge	1	0.3
7. Shale: dull-gray, chunky, strongly calcareous, finely arenaceous. Contains poorly preserved fossils. Weathers to angular fragments	4	1.2
6. Shale: olive-green, stained with tan limonite, soft, fissile, fossiliferous. Weathers to extremely tiny pale-gray fragments. Base of locality 2, zone 2	4	1.2
Note: Reeside identified the following species from this zone:		
Camptonectes bellistriatus Meek		
Boring sponge like *Cliona*		
Gryphaea aff. *G. planoconvexa* Whitfield		
Pachyteuthis densus (Meek and Hayden)		
5. Limestone, pale-gray, weathers straw-buff to tan, finely crystalline, siliceous, flaggy-bedded, fossiliferous. Breaks down to angular fragments and forms buff-tan, finely sandy soil. Contains tiny brachiopods and small pelecypods. Top of fossil locality 2, zone 1	23	7.0
4. Shale, gray, fissile, calcareous; buff-yellow and thick-bedded near base. Contains pelecypods	3	0.9

	Feet	Meters
3. Limestone: dull-gray, thin-bedded, hard, weathers to sharp angular fragments. Surface of some beds nearly covered with triturated brachiopods and pelecypods. Collection from this limestone	½-1	0.3
2. Shale and limestone: pale-gray, arenaceous, calcareous shale. Basal 1 foot is pale-gray, argillaceous and finely arenaceous limestone. Pelecypods common. Base of fossil locality 2, zone 1	10	3.0

Fossils:

Astarte packardi White
Camptonectes bellistriatus Meek
C. cf. *C. extenuatus* (Meek and Hayden)
Eumicrotis curta (Hall)
"Fucoids," apparently algal rather than trails or borings
Ostrea strigilecula White
Pleuromya weberensis (Meek and Hayden)
Protocardia cf. *P. shumardi* (Meek and Hayden)
Quenstedticeras (Eboraciceras) sp.
Volsella sp., probably new

Note: Reeside identified the species in the above list. Concerning them he says: "This fauna appears to me to be younger than those of the lower Ellis of Central Montana. I would place it as not older than upper Ellis (Divesian)."

1. Shales and limestone: dark-gray, fissile, soft shale, which grades up into pale-gray, finely arenaceous shale and, 16 inches from the top are 1½- to 3-inch beds of blue-gray, finely crystalline, nodular, pure limestone. Beds weather tan on joint faces. Ellis covered from this unit to base of formation	4	1.2
Total thickness of measured Ellis	244	74.2
Total thickness of measured section	384	116.9

MISSISSIPPIAN *(Hannan limestone)*

The contact of the Hannan and Ellis formations is exposed on the crest of the cliff of Hannan limestone which rises above the road on the west side of and approximately a quarter of a mile north of the mouth of Hannan Gulch. At this locality the basal beds of the Ellis formation consist of 6 to 8 feet of pale pink- and buff-gray, hard, dense, vitreous quartz.[2] The basal bed, as much as 4 feet thick in places, rests on the irregular and slightly channeled upper surface of the dolomite of the Hannan. The thickness of unexposed Ellis strata between the basal quartzite and the shale and limestone of No. 1 in the Lime Gulch section cannot be more than 10 to 15 feet. The total thickness of the Jurassic rocks, therefore, in the southern part of the Sawtooth Range is approximately 250 feet. Stebinger (1918, p. 155) gave 240 feet for the thickness of the Ellis east of the mountains in the southeastern part of the Saypo quadrangle.

Throughout the Sawtooth Range the soft Jurassic shales and limestones form gentle slopes, benches, or saddles below the ridges of Hannan

[2] The base of the Ellis in the Sweetgrass Hills (Sanderson, 1931, p. 30-31) is formed of similar dense arenaceous beds but is 30 feet thick.

limestone and weather to green-gray and buff-tan soil. The Ellis generally is poorly exposed in mountains because it weathers rapidly and also is much covered with talus of Hannan limestone or with glacial drift.

CRETACEOUS SYSTEM

KOOTENAI FORMATION

The Lower Cretaceous is represented in the Saypo quadrangle by the Kootenai formation (Dawson, 1885, p. 2) which rests almost conformably upon the Ellis. Stebinger (1918, p. 156) gave the thickness of the Kootenai in the plains area, on the eastern part of the quadrangle, as approximately 900 feet and described the rocks there as follows:

> "The Kootenai in the area here described . . . is composed essentially of shale and clay shale that are mostly red and purple, alternating with six to eight beds of gray to greenish-gray, medium to coarse grained sandstone between 30 and 60 feet in thickness and rather evenly distributed throughout the formation. The red color is characteristic of the Kootenai, making it readily identifiable in the field. The only other red beds in the section occur at a much higher horizon near the center of the Montana group."

In the southwest part of the Saypo quadrangle the Kootenai is exposed in the Sawtooth Range, in the valleys of the North and South Forks of Sun River, and along the lower eastern slopes of the Lewis and Clark Range just east of the Lewis overthrust. Throughout the southern part of the Sawtooth Range the soft Kootenai rocks have been eroded more rapidly than the Hannan limestone which stands as bold ridges above the valleys cut in the Jurassic and Cretaceous shales and sandstones (Pl. 3, fig. 2). The thickness of the Kootenai in the southwest part of the quadrangle is unknown but probably is approximately the same as in the plains area immediately east of the mountains. A section of the formation was not measured within the mountains because faults have repeated parts of the formation in many places, and glacial drift, soil, and talus cover the rocks to such an extent that a complete undisturbed section was not found.

Although the Kootenai is clearly demarcated from the underlying buff and gray, fossiliferous, calcareous shale and limestone of the Ellis, a sharp and well-defined contact between the two formations has not been seen in the Saypo area. Likewise, in the Sawtooth Range the upper part of the Kootenai is a sandstone and is overlain by the basal Blackleaf sandy member of the Colorado shale (Stebinger, 1918, p. 158-161). The contact between the Kootenai and Colorado was arbitrarily taken at the base of the first black shale above the youngest red-weathering sandstone.

The exact age in the southwest part of the Saypo quadrangle of the rocks assigned to the Kootenai formation cannot be determined precisely

upon the contained fossils because the exact species of plants known from the Kootenai in Alberta (Dawson, 1886, Pls. 1-2) and central Montana (Knowlton, 1907, Pls. 11-14) have not been found in the Sawtooth and Lewis and Clark ranges. In the upper part of the formation in the Sawtooth Range the writer found the following species which Reeside identified: *Unio* sp., *Corbicula?* sp., *Goniobasis? increbescens* Stanton, *G.? pealei* Stanton, and *Amnicola? cretacea* Stanton. Concerning this fauna Reeside (personal communication) says:

> "This fresh-water coquina contains forms most like those of the so-called Dakota of Yellowstone Park, now known to be Kootenai. It does not contain any of the usual nonmarine species that appear in the Bear River formation and the basal part of the Colorado. I would call the fauna Kootenai."

In the upper part of the red shales on Sheep Mountain fragments of a trunk or large limb of *Sequoia* sp. were found in 1940. Brown (personal communication) who identified the genus says: "Until the species of *Sequoia* wood are properly differentiated, little or nothing can be said about their significance as to geologic age."

The stratigraphic position of the Kootenai rocks in the area between the upper Ellis and basal Colorado marine shales, plus the lithologic similarity of the rocks with those in the Kootenai on the plains area contiguous to the Sawtooth Range, are, however, strong evidence that the rocks are of Kootenai age and are doubtless the western extension of that formation.

The Kootenai probably is not homogeneous throughout the southwestern part of the Saypo quadrangle, but the formation exhibits the following characteristics in several exposures examined in connection with geologic mapping. The basal 125 to 150 feet consists of tan and greenish-gray, thin- to thick-bedded, cross-bedded, fine- to medium-grained, calcareous sandstone, which contains numerous grains of hematite and limonite and weathers deep brown and buff tan.

The basal sandstone is overlain by an unknown thickness of maroon chunky to fissile shale and intercalated thin-bedded, soft, red and tan sandstone, which grades upward into brick-red, soft, friable sandstone and thin inter-bedded maroon shale.

The third conspicuous unit of the Kootenai consists of green-gray, coarse-grained, thick- and thin-bedded sandstone, which contains much triturated mica and weathers drab green tan. Green shale beds are interbedded with the sandstone. A unit 25 to 30 feet thick of interbedded conglomerate and sandstone occurrs 40 to 50 feet below the top of the green massive sandstone. The conglomerate is composed of subangular to rounded black chert and white and glassy quartz pebbles embedded in a matrix of arenaceous siliceous clay. The pebbles

range from an average of less than ¾ to as much as 1½ inches in diameter. The interbedded sandstone is green gray, medium-grained, and contains coarse grains of black chert, which causes a salt and pepper appearance.

Lenticular beds of limestone occur 50 to 70 feet above the conglomerate. In the southern part of the area one such lens, 4 to 7 feet thick, is dull- and light-gray, fossiliferous, thin- and nodular-bedded limestone and contains the species identified by Reeside and referred to above. North and west of this locality, east of the head of Neal Creek (Fig. 2), and north of Bench Mark Creek, limestones containing the same species are 30 to 50 feet or more thick and are black gray and thick-bedded. Some beds contain thin, curved flakes of calcite, which simulate fragments of pelecypod shells. Concerning this phenomenon Reeside (personal communication) says:

> "Material like that of this collection—thin sheets of calcite with a laminar structure but very irregular general form—has been attributed to the Algae by various writers. It is almost certainly not the shells of pelecypods. Almost identical material has been found in the Paleozoic and at various later levels."

The rocks which form the upper part of the Kootenai formation in the map area are tan-gray and pink, cross-bedded, hard, quartzitic, strongly banded sandstone, interbedded with thin-bedded, micaceous sandstone marked by worm borings.

In dealing with the Kootenai rocks in the Sawtooth and Lewis and Clark ranges it must be remembered that the rocks have been so intensely faulted and locally sheared and folded that an orderly fixed lithologic sequence is rarely recognizable.

COLORADO SHALE

General statement.—Of the Upper Cretaceous formations present in the eastern part of the Saypo quadrangle and underlying the plains east of the Sawtooth Range, only the Colorado shale is represented in the southwest part of the quadrangle in the Lewis and Clark and the Sawtooth ranges. The Upper Cretaceous formations in the plains area recognized by Stebinger (1918, p. 154) are the Colorado shale (with the basal Blackleaf sandy member), Virgelle sandstone, Two Medicine formation, Bearpaw shale, and Horsethief sandstone. In the plains region the Colorado is 1800 feet or more thick, and the upper four formations which comprise the Montana group in that area have a combined thickness of almost 3600 feet. All the formations probably were deposited in the area now occupied by the Sawtooth and Lewis and Clark ranges and must have been comparable in thickness to those east of the mountains. Two factors readily account for the absence

of most of the Upper Cretaceous rocks within the present mountains: (1) erosion since Eocene time, and (2) faulting that cut out most of the Upper Cretaceous rocks and thrust the Hannan, Ellis, and Kootenai so closely upon one another that space does not exist for the Upper Cretaceous sediments within the narrow fault blocks.

Lithology of Colorado shale in Sawtooth Range.—The only places the Colorado has been recognized are at the head of Beaver Creek Valley and in Patrick's Basin (Fig. 2). The Colorado shale probably is present also in the broad valley of the North Fork of Sun River but is covered there with alluvium and Pleistocene glacial drift.

At the head of Beaver Creek on the east side of the valley the lower part of the Colorado is exposed and rests on the upper sandstones of the Kootenai. The basal part of the Blackleaf sandy member, approximately 180 feet thick, consists of black-gray, slightly fissile shale, which contains occasional intercalated dark-gray, thin-bedded sandstones in the lower part and a 2-foot bed of black sandstone in the upper part of the unit. The shale becomes increasingly arenaceous toward the top of the unit. Brown and black-gray, extremely nodular-bedded sandstone, possibly 30 feet thick, rests on the basal black shale. The sandstone weathers red and black gray and breaks down to rounded boulders and nodular bedding surfaces. Above the sandstone is 350 feet or more of alternating units of black-gray fissile shale and rusty-tan-weathering sandstone. In the upper part of the unit is a zone 60 feet thick of black shale in which are intercalated a number of lenses 2 to 10 inches thick and as much as 60 feet long of very manganiferous, hard, dense shale. Within this zone most of the bedding surfaces of the shale are coated with pyrolusite or orange-colored limonite. At the top of the 350-foot unit is several feet of green-tan-weathering, micaceous, quartzitic, fossiliferous sandstone from which the following species, identified by Reeside, were collected: *Lingula* sp., *Nuculana* sp., *Ostrea* sp., *Thracia?* sp., *Cardium* n. sp. ("like an unnamed species in the Bear River and Aspen formations"), *Corbula pyriformis* Meek var., *Corbula* cf. *C. nematophora* Meek, and *Pyrgulifera humerosa* Meek var. Concerning this faunule Reeside (personal communication) says: "This assemblage of species belongs to a fauna now known at a number of localities near the base of the Colorado group—specifically in the Bear River and Aspen formations."

Black fissile shale and widely separated thin and irregularly distributed beds of sandstone characterize the Colorado shale above the Blackleaf sandy member in the southern part of the Sawtooth Range. These rocks have been seen only in isolated exposures in Patrick's Basin west

of Beaver Creek. In the southwestern part of the Saypo quadrangle the boundary between the Colorado and Kootenai is rarely exposed.

QUATERNARY SYSTEM

GENERAL STATEMENT

The only Cenozoic deposits in the southwestern part of the quadrangle are unconsolidated Pleistocene till and fluvio-glacial sands and gravels, and Recent alluvium laid down since the retreat of the alpine glaciers of late Wisconsin stage.

PLEISTOCENE DEPOSITS

The Lewis and Clark and the Sawtooth ranges were glaciated at least twice during the Pleistocene. Sufficient field work has not been done to date the early stage whose presence is indicated by erratics of Belt quartzite and sandstone from the Lewis and Clark Range and by diorite erratics (Pl. 2, fig. 2) which are present on the crests of the ridges in the Sawtooth Range above 7000 feet in altitude. Old eroded cirques with maturely developed talus slopes and old moraines dissected by the latest glaciers also indicate the older stage of glaciation. The absence of leaching or extreme weathering of the older moraines would seem to indicate that the first recorded Pleistocene stage is not older than the Illinoian or Iowan stage (Alden, 1932, p. 69-70).

During the Wisconsin stage most of the intermontane valleys were partly filled with ice, and numerous cirques and arêtes of the higher peaks were cut (Pl. 2, fig. 1). The ice in the Sawtooth Range found exit in the southern part of the Saypo quadrangle from the mountains eastward onto the plains through the Canyon of the Sun River and deposited the large moraine in front of the mountains in the southeastern part of the Saypo area (Alden, 1932, p. 120-121). Within the mountains, the valley glaciers deposited lateral, medial, and recessional moraines on the sides and bottoms of the valleys, and the waters from the melting ice spread local outwash plains on the bottoms of the wider valleys. The valleys of the streams flowing east from the Lewis and Clark Range, the divides between the streams, and much of the broad valley of the North Fork of Sun River are covered with thick drift which blankets most of the older formations in the area.

RECENT ALLUVIUM

The youngest deposits in the map area have been laid down by streams since the retreat of the Wisconsin ice. These deposits are muds, sands, and gravels brought down from the higher levels by tributary streams to the more mature valleys where they are spread out as thin flood plains. In some of the valleys beaver dams have produced temporary small lakes

in which fine silts have collected. With few exceptions, such as the North Fork of Sun River and Beaver Creek, most of the valleys contain little alluvium as the streams are actively cutting downward and producing narrow canyons in the bottoms of the U-shaped valleys cut by the Wisconsin glaciers.

HIATUSES WITHIN MAP AREA

GENERAL STATEMENT

Five hiatuses are recorded in the stratigraphic section within the Saypo quadrangle, and each is of such magnitude that the breaks in sedimentation are readily recognized in the field. Angular discordances of the strata between any two of the formations, however, are not apparent within the quadrangle, and the duration of each hiatus is established upon other criteria. The five major breaks in sedimentation took place between: (1) Cambrian and pre-Cambrian, (2) Devonian and Cambrian, (3) Mississippian and Devonian, (4) Jurassic and Mississippian, and (5) Pleistocene and Cretaceous.

CAMBRIAN—PRE-CAMBRIAN UNCONFORMITY

In 1935 Deiss (1935, p. 95-124) described the unconformity in western Montana between the Belt series and Middle Cambrian and gave evidence which indicated that erosion removed at least 20,000 feet of upper Belt rocks before the Cambrian sea invaded the area. Throughout most of the Lewis and Clark Range the basal Middle Cambrian Flathead sandstone rests in apparent conformity on the underlying Belt rocks. In spite of the apparent conformability of the two series, the Flathead sandstone rests on different Belt strata at different places in the range, and, locally, the angular relationships of the beds on the two sides of the erosion surface are discernible (Deiss, 1939, Pl. 11, fig. 1).

The relationships of the Belt and Cambrian along a section from Pentagon Mountain in the northern part of the Lewis and Clark Range to the Three Forks quadrangle, Montana, 225 miles southeast, were described in 1935. Part of that section between Prairie Reef (Silvertip quadrangle) and Ford Creek (Coopers Lake quadrangle) passes through the southwestern part of the Saypo quadrangle. In discussing this part of the section Deiss (1935, p. 112) said:

> "The northern boundary of the region, which was more greatly elevated in northwestern Montana, lies between Prairie Reef and Ford Creek. These two localities are 15 miles apart, and within this distance the Missoula group was eroded at least 5,500 feet. The evidence for the uplift shown in this part of the diagrammatic section (Pl. 8, fig. 4), is that at Prairie Reef the Flathead lies upon Missoula group argillites approximately 10,000 feet stratigraphically above the Helena limestone, and at Ford Creek it lies upon fissile argillites of the same group, which are only 4,500 feet above the Helena."

At Hoadley Reef the Flathead rests upon the green argillite in the upper part of the Ahorn formation. This horizon is estimated to be 7300 feet stratigraphically above the Helena limestone. If this estimate is correct and if the rocks equivalent to the Missoula group were 16,000 feet or more thick in the Lewis and Clark Range, as they are in the Swan Range (Deiss, 1935, p. 106, Pl. 7) at least 8700 feet of Belt rocks was eroded from the southwestern part of the Saypo quadrangle between the cessation of Belt deposition and early Middle Cambrian. If Belt sedimentation continued in the area nearly until the beginning of the Cambrian, then most of the erosion must have been accomplished during the Waucoban (Lower Cambrian).

The oldest Cambrian rocks in the Saypo quadrangle (Flathead sandstone) probably are approximately equivalent in age to the *Albertella* zone in British Columbia and Alberta (Deiss, 1940, p. 782-783). Sediments containing species of the basal Middle Cambrian fauna, *Kochaspis liliana* (Walcott), are unknown in northwestern Montana. Therefore, the unconformity in the area can be dated as post-Belt and pre-Middle Cambrian.

DEVONIAN—CAMBRIAN HIATUS

In the southwestern part of the Saypo quadrangle the youngest known Cambrian species belong to the *Cedaria* fauna (early Upper Cambrian). Above the beds which contain *Cedaria* on Prairie Reef is 350 feet of unfossiliferous white-gray dolomite and drab-green shale (Devils Glen, and upper part of Switchback). The Devils Glen is overlain almost conformably by unfossiliferous buff dolomites and shales which are transitional upward into limestone that contains species of *Atrypa, Spirifer, Favosites,* and others, which Cooper (personal communication) considers characteristic of Upper Devonian. Fossils of the Ordovician, Silurian, and Lower and Middle Devonian have not been found in northwestern Montana, and other evidence is unknown which would indicate that seas were present in the area during these periods and epochs.

The contact between the Devils Glen and Devonian strata is apparently conformable in the Saypo area, but in the northeast corner of the Silvertip quadrangle at Pentagon Mountain an irregular erosion surface is exposed between the strata of the two systems (Deiss, 1939, Pl. 11, fig. 2). The Devils Glen was eroded approximately 385 feet between Pentagon and Monitor mountains in the Lewis and Clark Range (Deiss, 1939, p. 50). The summary of the conditions between Upper Cambrian and Upper Devonian in northwestern Montana applies well to the southwestern part of the Saypo quadrangle (Deiss, 1939, p. 50-51):

"... the northwestern part of the Cordilleran trough within the United States was a lowland, and therefore, only slightly eroded during Upper Cambrian, Ordovi-

cian, Silurian, and Lower [and Middle] Devonian time. If this be true, the region did not experience any orogenic movements between the Cambrian and Devonian. Instead, the only movement indicated at the end of the Middle [Upper] Cambrian is gentle tilting from the Montana Island[2] southward and southeastward. Possibly the greater amount of erosion (386 feet, 117.8 m.) was accomplished by the end of the Upper Cambrian, the region lying almost at sea level and therefore undergoing little degradation throughout the Ordovician, Silurian, and Lower [and Middle] Devonian. Consequently, the hiatus in northwestern Montana between the top of the Devils Glen dolomite and the superjacent White Ridge limestone (Fig. 4) seems to represent a long interval of nondeposition."

MISSISSIPPIAN—DEVONIAN HIATUS

Within the southwestern part of the Saypo quadrangle the Hannan limestone rests conformably on the upper beds of the Devonian, and were it not for the conspicuous lithology of the conglomerate at the base of the Hannan limestone the two systems would appear to represent continuous deposition. The basal Mississippian, however, rests on strata lower in the Devonian section in the Sawtooth Range than in the Flathead and parts of the Lewis and Clark ranges to the west. At different places in these ranges Mississippian limestone lies on Devonian beds as much as 425 feet apart stratigraphically. Further, the pebbles in the basal conglomerate of the Mississippian (Silvertip) consist of younger Devonian limestones in some sections and older Devonian in others. From these facts the writer (1933, p. 45) concluded that:

"The area was elevated . . . in late Jefferson time, while deposition continued in the northern and southern portions, resulting in the sediments subsequently indurated to form the Spotted Bear member (youngest Devonian). At the end of Spotted Bear time the entire region was raised only slightly and both the then newly deposited Spotted Bear limestone, as well as the underlying Lone Butte limestone, underwent but little erosion. . . . The basal Madison limestone-conglomerate was deposited in apparent conformity (but in different parts of the region) upon the upper and but slightly eroded surfaces of both the Spotted Bear and the underlying Lone Butte members. By using the thickness of the Spotted Bear limestone in the Dearborn section, disregarding the thickness eroded during the Devonian—Mississippian interval, it may be determined that the difference in thickness of the limestones upon which the basal Madison beds rest, and hence the amount of disconformity between the beds of the two systems, is approximately 425."

The amount of erosion, if any, which took place within the Saypo quadrangle between the Mississippian and Devonian has not been determined.

JURASSIC—MISSISSIPPIAN HIATUS

The break in sedimentation between the Hannan and Ellis formations on the east side of the Saypo quadrangle was noted by Stebinger (1918, p. 156) who said:

"The shales of the Ellis, so far as the evidence in this area alone indicates, lie in apparent conformity on the Mississippian limestone, there being no indications of angular discordance or a period of erosion between the deposition of the Mississippian and that of the Ellis. This conformity is only apparent, however, for at several localities within a radius of 100 miles rocks of upper Carboniferous and

[2] Now known as the positive element Montania (Deiss, 1941, p. 1095-1097).

probably Triassic age, aggregating as much as 2,000 feet in thickness, are known to be present in the interval between the lower Carboniferous limestone and the Jurassic of this general region."

In 1927 Collier (1927, p. 130-131) discussed the unconformity between the Madison and Ellis formations in the plains area of central and northern Montana and southern Alberta but did not discuss the break in the front ranges of the Rockies. Collier (1927, p. 131) concluded that:

"This unconformity therefore represents a great peneplain eroded in late Triassic or early Jurassic time which extended about 800 miles north and south and 200 or more miles east and west. It was eroded on Triassic rocks at its south end; on Quadrant and Madison rocks in its middle part and on Madison and Devonian rocks at its north end."

In the southern part of the Sawtooth Range the contact of the Ellis and Hannan formations is well exposed only on the west side of Hannan Gulch (*see* discussion following Lime Gulch section) where the basal sandstone of the Ellis rests on the upper dolomite of the Hannan. The erosion surface at the top of the Hannan is only slightly irregular. The wide and undulating channels in the upper surface of the dolomite are 4 to 10 inches deep, and were filled with basal sands of the Ellis.

If the upper part of the Hannan limestone is equivalent to the Brazer in southeastern Idaho, which Girty (Mansfield, 1927, p. 69, 71) assigned to the early Chester and doubtfully to the Meramec, and if the Ellis in the map area is Divesian and Argovian in age, the southern part of the Saypo quadrangle was above sea level throughout the late Chester, all of the Pennsylvanian, Permian, and Triassic, and all of Lower, Middle, and part of Upper Jurassic. The absence of sediments and fossils representative of these epochs and periods in northwestern Montana indicates that Montania was again a positive element in the Cordilleran region as it was until Middle Cambrian and also from the end of the Upper Cambrian until Upper Devonian.

TERTIARY EROSION

The Colorado shale and Montana group of Upper Cretaceous formations probably were deposited in the southwestern part of the Saypo quadrangle as in the area occupied by the plains east of the Sawtooth Range. Whether or not the St. Mary River formation was deposited in the southwestern part of the quadrangle is doubtful. After the Laramide Revolution the newly elevated mountains must have been capped with Upper Cretaceous sediments. Throughout the Tertiary and much of the Pleistocene the Upper Cretaceous rocks in the mountains were eroded and the debris transported eastward onto the plains. Consequently, by the end of the Tertiary not only the Upper Cretaceous but much of the Kootenai and Ellis sediments were removed from the southwestern part

of the Saypo quadrangle. Since the withdrawal of the Horsethief Sea, late in the Cretaceous, the area has not been flooded with marine waters and, unless local Tertiary lakes formed in the mountains, younger strata were not deposited. The only post-Cretaceous sediments laid down in the southwestern part of the Saypo area are, therefore, unconsolidated glacial drift and post-Pleistocene alluvium.

IGNEOUS ROCKS

TYPE AND OCCURRENCE

The igneous rocks in the Saypo quadrangle are medium to finely crystalline diorite and gabbro sills. Extrusive rocks have not been found in the mountains in the southwestern part of the quadrangle, but in the eastern part of the Coopers Lake quadrangle (Fig. 1) just southeast of the Saypo area is an eroded volcanic plug or neck, and between the Dearborn and Missouri rivers east of the Coopers Lake quadrangle are extensive beds of fragmental volcanics (Clapp, 1932, p. 28). Volcanic ash and possibly flows may have covered parts of the map area in the late Cretaceous or early Tertiary, but if so they were eroded before the Pleistocene.

The sills are exposed only in the western third of the Saypo area, west of the meridian 112°50′, except near the southern edge of the quadrangle in Patrick's Basin and Renshaw Creek (Fig. 2) where they extend 1½ miles east of the meridian. The sills are intruded into the Kootenai and possibly into the overlying shales and sandstones of the Colorado along the west side of the Sawtooth Range (Fig. 3, C-C′, middle part of section). In this eastern belt at least three sills are folded (Pl. 4, fig. 1) and eroded so that they form nine exposures of diorite in the upper part of Patrick's Basin. Other sills comprise a western belt in the Saypo area where they are intruded into the argillite and quartzitic sandstone of the Hoadley along the east flank of the Lewis and Clark Range. Strangely, the Paleozoic rocks in the Lewis and Clark and the Sawtooth ranges do not contain sills or other intrusions, a condition characteristic of all northwestern Montana (C. H. Clapp, 1932, p. 27). Another unexplained fact is the absence of sills in the Mesozoic formations east of meridian 112°50′ in the Saypo quadrangle (Fig. 3).

In the eastern belt, along the west side of the Sawtooth Range, the sills are more numerous and continuous than in the western belt and are folded into open anticlines (Pl. 4, fig. 2) and synclines whose axes in the southern part of the area strike N. 40° W. North of the Sun River (east of the North Fork Valley) the trend is almost due north. In both belts the sills are intermittent, a characteristic which may be partly apparent where the sills are dissected by streams that flow across the strike and whose valleys

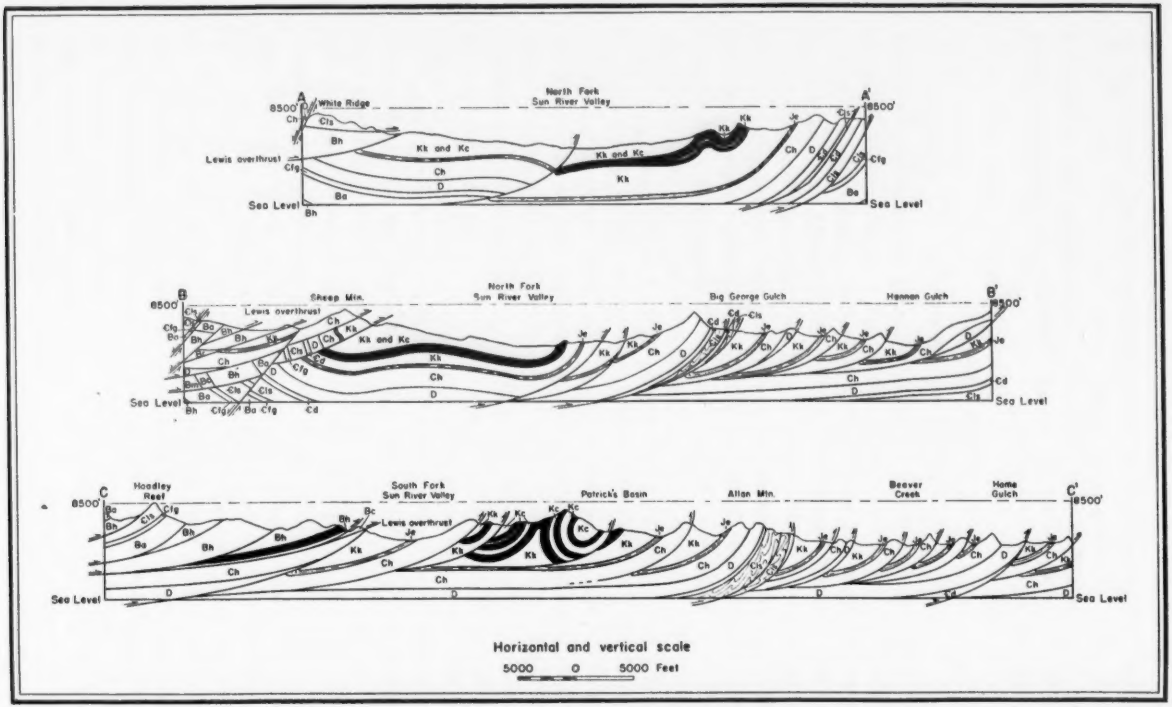


FIGURE 3.—*Structure sections across southwest part of Saypo quadrangle*

(*See* Figure 2.) Kc = Colorado shale, Kk = Kootenai formation, Je = Ellis formation, Ch = Hannan limestone, D = Devonian (formations unnamed), Cd = Devils Glen dolomite, Cls = Cambrian limestones undifferentiated, Cfg = Flathead sandstone and Gordon shale combined, Ba = Ahorn quartzite, Bh = Headley formation, Bc = Cayuse limestone, Bm = Miller Peak argillite. Solid black = late Cretaceous diorite sills.

are covered with glacial drift. The sills high on the slopes of bedrock, however, pinch out in some places and reappear in others. This abrupt lateral termination of one of the sills is well shown on the north side of the valley south of Deer Creek, and also on the northeast spur of Dead Man Hill northwest of Bench Mark 6865 (feet). The condition also occurs at several points in Patrick's Basin.

COMPOSITION OF SILLS

Most of the sills in the Kootenai and Colorado rocks exhibit a holocrystalline medial phase and a phaneritic or aphanitic lateral or contact phase. The middle part of the sills is medium to coarsely crystalline diorite which locally contains crystals of free quartz. Ross (personal communication) who studied specimens of this rock says: "Contains a little quartz but hardly enough to justify the term granodiorite. Very few primary ferromagnesian minerals. These now completely altered to calcite. Secondary minerals, calcite and interstitial biotite, partly altered to chlorite." In the contact zone the rock is finely crystalline to aplitic and in some outcrops is porphyritic and contains phenocrysts of lathlike glassy plagioclase as much as a quarter of an inch long. Leucoxene is abundant locally, and the feldspar is altered to kaolinite and the pyrite to limonite. The lack of intense metamorphism of the confining sandstones and the narrowness of the altered zone of country rock suggests that the magma was fairly dry. The diorite is cut by aplite dikes which range from ¼ to 3 inches in thickness in many localities. The basic magma apparently cooled and cracked, and much later an acidic magma filled the fissures. The diorite weathers rusty brown stained red and forms peaks and ridges above the surrounding Kootenai and Colorado shales and sandstones (Pl. 4).

The sills intruded into the Hoadley formation vary in composition more than do those in the Mesozoic formations and may be generally more basic. The sills in the western belt consist of coarsely crystalline gabbro in the medial part and of much finer-grained darker rock of the same type near the contact where actinolite locally is developed. Concerning these rocks Ross (personal communication) says: "Augite forms nearly half the rock. Magnetite or ilmenite is abundant. A few grains of biotite are present." In other places the sills in the Hoadley show small magmatically segregated masses of large crystals of pink orthoclase and smaller ones of hornblende, biotite, and quartz. Some of the masses are lenses as much as 8 inches long and 1½ inches wide. Smaller aplitic lenses of similar composition also are developed. Ross identified specimens of these lenses as quartz diorite and comments on them as follows:

Bh = Hoadley formation. Bc = Cayuse limestone, Bha = Miller Peak argillite. Cfg = Flathead sandstone and Gordon shale combined. Bu = undifferentiated. Bm = Mission formation (formations unnamed). Solid black = late Cretaceous diorite sills.

"Abundant biotite partly or completely altered to chlorite. Quartz forms an abundant micrographic intergrowth with feldspar. Augite is partly altered to hornblende and chlorite. A little calcite present." The texture of the medial part of the sills in the Hoadley is slightly coarser than that of the sills in the Kootenai and Colorado.

DEFORMATION OF SILLS

Numerous gash or tear fractures (Pl. 4, fig. 2) in the sills on the crests of anticlines indicate clearly that the intruded magmas hardened before deformation. Most of the open anticlines and synclines into which the sills were folded are so far from the faults in the region (Fig. 3, A-A′ and C-C′) that the folds cannot be attributed to drag along the thrust faults. These two facts suggest that the magmas were intruded before the earliest faulting in the area, probably late in the Cretaceous, and that the folding of the sills may have preceded the faulting.

STRUCTURAL PROVINCES IN SOUTHWEST SAYPO QUADRANGLE

The southwestern part of the Saypo quadrangle (Fig. 2) contains two sharply demarcated structural provinces which are separated by the southern extension of the Lewis overthrust (Willis, 1902, p. 331-333). The eastern province is largely in the southern part of the Sawtooth Range but extends westward to the South Fork of Sun River and also approximately 4 miles west of the North Fork of Sun River. The eastern province is characterized by high-angle, strike, dip-slip, reverse faults which trend N. 5° to 20° W. in the area south of the Sun River, and due north to N. 7° E. in the mountains north of the river. The western part of the eastern province contains the belt of folded sills in the Mesozoic rocks, and near its western edge contains two thrust faults of lower dip whose strikes average N. 40° W. The formations in the southern part of the Sawtooth Range dip steeply west and, except for the shales of the Cambrian and Kootenai at several localities, are not much folded (Fig. 3).

The western structural province is characterized by low-angle, strike and oblique, dip-slip, overthrust faults including the Lewis overthrust at the east edge of the province (Fig. 3, C-C′, west end of section). These faults are of greater magnitude than those in the eastern province and displace young Belt and old Paleozoic rocks which dip less steeply than the younger formations in the eastern province. The low-angle normal fault between Glenn and Moose creeks (Fig. 2) and the high-angle normal faults near White Ridge (Fig. 3, A-A′, west end of section) are also in the western province.

FIGURE 1. DRAG-FOLD AND FAULT IN KOOTENAI FORMATION
Near upper falls of Lange Creek.

FIGURE 2. HANNAN LIMESTONE THRUST UPON SHALE OF THE KOOTENAI FORMATION
West side near head of Home Gulch.
H = Hannan, K = Kootenai.

HIGH-ANGLE THRUSTS

FIGURE 1. SHARP FOLDS IN SILLS, AND HANNAN LIMESTONE THRUST UPON SHALE OF THE KOOTENAI FORMATION
West side of Patrick's Basin, looking northwest.
D = diorite sills, H = Hannan, K = Kootenai.

FIGURE 2. TENSION FRACTURES IN DIORITE SILL ON CREST OF FOLD
Peak unnamed but carries Bench Mark 6968. Looking northwest across valley of South Fork of Sun River.
D = diorite sills, K = Kootenai formation.

FOLDED DIORITE SILLS

FAULTS

GENERAL STATEMENT

Four types of faults are recognized in the two structural provinces: (1) high-angle thrusts, (2) low-angle overthrusts including the Lewis overthrust, (3) low-angle normal fault, and (4) high-angle normal faults.

HIGH-ANGLE THRUST FAULTS

The oldest faults in the southwestern part of the Saypo quadrangle are in the eastern structural province. These faults all dip steeply to the west at angles which range from 30° to 75° or more and average 55°. The faults exhibit slips of less than 1000 to more than 7000 feet and average 3000 feet. The throw ranges from less than 1000 to more than 5000 feet and may average 2000 feet. The faults bow slightly southwest. The convex or outside of the bow on the west is the upthrown, and the concave side on the east is downthrown. The strike of the faults is nearly parallel with the strike of the strata (Pl. 3, fig. 2), and the movement was nearly parallel with the dip of the fault surfaces. These strike, dip-slip, high-angle thrusts are characteristic of the eastern province and are responsible for the present position of the ridges and valleys and the geologic structure in the southern part of the Sawtooth Range (Fig. 3, B-B′ and C-C′, east of Sun River Valley).

The faults in the Sawtooth Range generally thrust Hannan limestone upon shales and sandstones of the Kootenai (Pl. 3, fig. 2), but in the south-central part of the Sawtooth Range Devonian and Cambrian limestones were thrust over the Kootenai (Fig. 3, B-B′, at Big George Gulch; C-C′, east of Allan Mountain). On the east side of Sawtooth Ridge, the high-angle thrust and its northern extension east of Castle Reef also threw Devonian against the Kootenai. Throughout most of the southern part of the range the dip of the strata is nearly that of the faults at the surface of the earth, but both the strata and fault surfaces must flatten downward. As a result of these relationships the Sawtooth Range is predominantly a series of steeply dipping, generally post-Cambrian, Paleozoic, and Mesozoic sediments repeated by a number of high-angle strike thrust faults.

The fault surfaces have been seen in several outcrops where the under surface of the overriding Mississippian limestone is curved but rarely is brecciated, and the shale and sandstone of the Kootenai beneath the fault surfaces is drag-folded. Gouge and conspicuous dragfolds were not produced in most places. On the high-angle thrusts confined to the Kootenai rocks, as in Lange Creek Valley at the upper falls, slickensides and drag occasionally are prominent (Pl. 3, fig. 1).

West of Patrick's Basin (Fig. 3, C-C′, west-central part of section) the dip of the thrusts decreases, and the trend of the faults is more west of north. Two such thrusts are marked by the ridges of Hannan limestone which cross the South and West Forks of Sun River in the southwestern part of the Saypo quadrangle. The western thrust is truncated by the Lewis overthrust, which south of Prairie Creek threw limestone of the Cayuse (Belt) on limestone of the Hannan (Mississippian), and north of Prairie Creek (Fig. 2) threw sandstone of the Hoadley (Belt) on sandstone of the Kootenai (Cretaceous). These facts indicate that the high-angle thrusts are older than the low-angle Lewis overthrust.

LOW-ANGLE OVERTHRUST FAULTS

The largest and most important low-angle overthrust in the map area is the Lewis overthrust, which marks the east edge of the western structural province and the boundary between the Sawtooth and the Lewis and Clark ranges in the southern part of the Saypo quadrangle. Southwest of the Lewis overthrust four similar but smaller overthrusts (Fig. 3, C-C′, west end of section) are confined to the Belt rocks, with the exception of the westernmost fault which threw upper sandstone of the Hoadley upon Devils Glen dolomite (Cambrian) (Pl. 5, fig. 2). The low-angle overthrusts have an average dip of 30° to the west. The dip of the faults is so nearly the same as the dip of the beds that in places they appear as bedding faults. The strike of the faults is nearly parallel with the strike of the beds, and the movement apparently was parallel with the dip of the fault surfaces. The amount of displacement along the faults is unknown, but as the overthrust on the west side of Hoadley Reef threw beds approximately 700 feet below the top of the Hoadley formation on the upper beds of the Devils Glen, the slip must have been at least 12,000 feet and the throw 5000 feet. The movement along the other overthrusts (except the Lewis) was less than that on the west side of Hoadley Reef.

Clapp (1934) estimated that in the Coopers Lake quadrangle just south of the Saypo area (Fig. 1) the Lewis overthrust has an average dip of 18°, a stratigraphic displacement of 5000 to 15,000 feet, and a slip of at least 5 or 6 miles. Just west of the South Fork of Sun River in the Saypo area the Lewis overthrust threw argillite of the Miller Peak on limestone of the Hannan, and thus at this point the fault has a stratigraphic displacement of approximately 12,000 feet and a slip of as much as 7 miles. The surfaces of the low-angle overthrusts have not been seen, but the strata adjacent to the faults are usually slightly drag-folded.

Along the east flank of the Lewis and Clark Range, from the south edge of the Saypo quadrangle north to Prairie Creek, the western structural province is characterized by these low-angle overthrusts which repeat parts of the Belt formations at least five times.

LOW-ANGLE NORMAL FAULT

The only low-angle normal fault recognized in the map area is east of White Ridge between Glenn and Moose creeks (Fig. 2). The Middle Cambrian limestones above the Gordon shale were faulted eastward against the Hoadley formation, and the Ahorn, Flathead, and Gordon formations were cut out (Fig. 3, A-A′, west end of section). The fault surface is covered with soil and a thick mantle of drift. Its strike and dip, therefore, are unknown, but by scaling the contact drawn on the map between the Cambrian limestones and the sandstone of the Hoadley, a dip of 5° to 7° is suggested.

This low-angle normal fault may have been formed after the low-angle overthrusts were completed and the lateral stresses subsided. The normal fault preceded the high-angle normal faults in the same area, however, because the latter cut the low-angle normal fault near the head of Glenn Creek (Fig. 3), and also just west of the Saypo quadrangle on the south side of Moose Creek. The mechanics which produced this structure are not yet understood, and additional field work must be done on the area.

HIGH-ANGLE NORMAL FAULTS

The youngest faults in the area probably occur in the western structural province between the upper parts of Prairie and Moose creeks (Fig. 2). In this area four normal faults dip 45° to 70° west and are bowed slightly to the northeast. The west or inside of the bow is the downthrown side of the fault. These high-angle normal faults are also dip-slip, and their strike is more or less parallel with that of the strata. Neither dragfolds nor gouge was observed along the fault lines. The faults are conspicuous in the field (Pl. 5, fig. 1) because they dropped rocks of different color or composition against each other. For example, on the ridge north of the North Fork of Bear Creek the Flathead sandstone and Ahorn quartzite were dropped against sandstone of the Hoadley, and near the heads of Bear and Glenn creeks the Cambrian limestones were dropped against the Ahorn and Hoadley formations respectively. Likewise, near the crest of White Ridge the Hannan was dropped against the Devonian which in turn rests against Steamboat and Pagoda limestones.

The high-angle normal faults truncate both the low-angle normal fault and the low-angle overthrusts (Fig. 3, B-B′, west end of section)

and, therefore, are the youngest faults in the region. They are believed to have formed after the lateral stresses which produced the thrust faults had relaxed and the rocks were settling to the positions they finally attained after the Laramide orogeny. Evidence was not seen which indicates that such faulting took place in the Saypo quadrangle during the late and post-Tertiary uplift of the Rocky Mountains.

FOLDS

GENERAL STATEMENT

Folds are the minor geologic structures in the southwestern part of the Saypo quadrangle and are largely confined to the Cambrian and Cretaceous formations and to the sills in the Mesozoic sediments. The rocks are not folded strongly except in two rather sharply defined belts. The soft shales and sandstones of the Colorado and Kootenai may be much drag-folded, however, along the high-angle faults (Pl. 4, fig. 1) but are covered in most places with glacial drift to such an extent that minor structures are rarely visible.

EASTERN BELT OF FOLDING

The eastern belt of folded rocks is one-half to three-quarters of a mile wide and extends N. 15° W. from the West Fork of Beaver Creek along the east side of Allan Mountain to the Sun River, continuing almost due north nearly 5 miles along the west side of Big George Gulch (Fig. 2). In this belt the shales and thin-bedded limestones of the Steamboat and Pagoda formations were dragged into a closed anticline and syncline south of the Sun River, and into a series of tightly squeezed recumbent and faulted anticlines and synclines north of the river. The closed folds probably are minor flexures in the competent beds which were deformed when the rocks were thrust eastward along the largest high-angle thrust fault in the southern part of the Sawtooth Range (Fig. 3, B-B′, at Big George Gulch; C-C′, east of Allan Mountain). The crest and most of the east limb of the large anticline have been removed by erosion.

WESTERN BELT OF FOLDING

The western belt of folding is irregularly V-shaped and extends from the south edge of the quadrangle northwest to the South Fork of Sun River, then north beyond Cabin Creek. At the southern end the belt is almost 2 miles wide and at the northern end near Cabin Creek the belt may be 6 miles wide. The eastern boundary of the belt is formed by the slices of Hannan limestone on the east side of Lange Creek and, north of Sun River, by the fault slice of Hannan limestone on the west side of the Sawtooth Range. The western boundary is formed by the

FIGURE 1. CAMBRIAN LIMESTONES DROPPED AGAINST BELT ARGILLITES AND QUARTZITES
Looking north across Moose Creek Valley along east side of Lewis and Clark Range. C = Cambrian limestones, Bh = Hoadley formation.

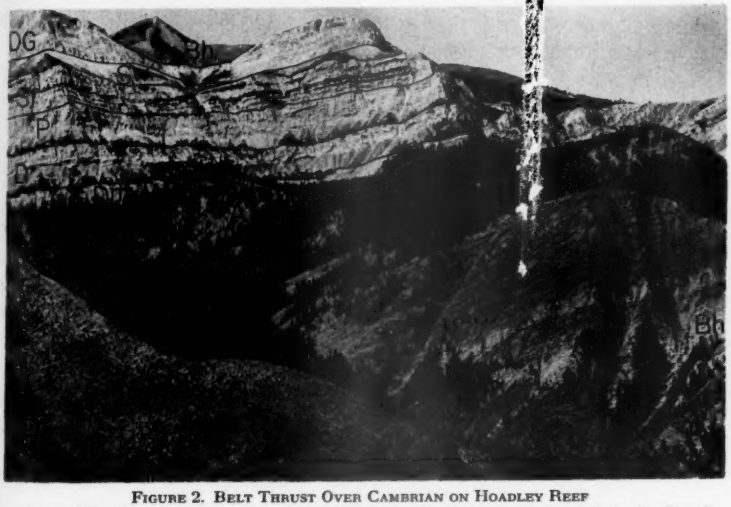


FIGURE 2. BELT THRUST OVER CAMBRIAN ON HOADLEY REEF
A = Ahorn, Bh = Hoadley (Belt); DG = Devils Glen, Sw = Switchback, S = Steamboat, P = Pagoda, D = Dearborn, and Dm = Damnation, formations.

NORMAL AND OVERTHRUST FAULTS

fault slice of Hannan limestone (Pl. 4, fig. 1) whose north end forms Sheep Mountain.

The most prominent folds in the area are in the western belt and are all in the Mesozoic sediments and particularly in the sills (Pl. 4, fig. 2). In this belt the sills were the most competent beds and were deformed into a continuous series of open asymmetrical anticlines and synclines (Fig. 3, C-C′, central part of section) which plunge approximately 8° northwest and strike N. 35° to 40° W. Three or four such folds are exposed in the southern part of the belt. The folds near Circle Creek on the northeast side of the belt (Fig. 3, A-A′, east-central part of section) strike nearly N. 15° W. and plunge more steeply than those in the south. In places, such as the head of Patrick's Basin, the folds are cut by high-angle thrust faults (Pl. 4, fig. 1). Throughout most of the belt, however, the folds are open and are not associated intimately with faults. Further, the tension fractures in the sills on the crests of the anticlines (Pl. 4, fig. 2) indicate that the sills were deformed after they had cooled and probably as the result of lateral stresses, not as drag along faults. The folds in the western belt may have been produced early in the Laramide orogeny before the high-angle thrust faults. If so the folding of the sills and Mesozoic sediments in the western belt may represent the first stage in the structural history of the area. Although the cause is not clear, both the eastern and western belts of folding are within the central and western parts of the eastern structural province in the southwestern part of the Saypo quadrangle.

SEQUENCE OF GEOLOGIC EVENTS

PROTEROZOIC TO CENOZOIC SEQUENCE

The part of the Cordilleran geosyncline within the Saypo quadrangle experienced a long period of quiet sedimentation which alternated with long and short periods of emergence when the area was only slightly above sea level. Marked angular unconformities are not recorded between any two formations within the area.

During the late Proterozoic the Saypo quadrangle was part of the newly formed Cordilleran geosyncline which, during Belt time, received more than 50,000 feet of clastic and calcareous sediments (Clapp, 1932, p. 22). More than 8000 feet of the upper Belt rocks (Miller Peak, Cayuse, Hoadley, and Ahorn formations) are exposed in the Lewis and Clarke Range in the southwest corner of the Saypo area.

Following the cessation of Belt sedimentation the rocks were elevated and tilted slightly and, during the Lipalian interval and the Lower Cambrian, approximately 8700 feet of the rocks was eroded.

Marine waters entered the region early in the Middle Cambrian, and in them were deposited 1800 to 1900 feet of limestone and shale. The region was elevated sufficiently to drain off the sea near the close of the period, but deposition continued in the Big Belt Mountains southeast of the Saypo area. The Saypo region remained slightly above sea level until Upper Devonian but did not suffer greatly from erosion.

During the Upper Devonian, marine waters covered the Saypo area until the close of the period. In the sea was deposited 1000 feet or more of calcareous and argillaceous sediments. At the close of the Devonian the area was again slightly elevated bodily but remained above sea level only until early in the Mississippian.

The last invasion of the region by marine waters in the Paleozoic occurred early in the Mississippian, when fine muds and calcareous oozes were deposited almost conformably upon limestone of the Devonian. This sea remained throughout much of the Mississippian and received the sediments which later were hardened into the Hannan limestone. The Saypo region was re-elevated just above sea level at the end of the Mississippian and remained above water until late in the Jurassic. Throughout this long time, however, very little erosion seems to have taken place because the Hannan limestone is uniform in thickness and its upper surface was not deeply channeled.

The Upper Jurassic in this part of the Cordilleran trough was a time of quiet sedimentation. The surrounding lands, however, must have been temporarily more elevated than during the Devonian and Mississippian as the Ellis formation contains more mud and sand than do the earlier formations. The sea probably was drained from the region before the close of the Jurassic as fossils younger than Argovian (*Cardioceras* zone) are unknown from the Jurassic rocks in northwestern Montana.

Before the beginning of the Lower Cretaceous the local geography changed considerably. The shallow seas and low lands of late Jurassic were replaced by temporary lakes, subaerial basins, and flood plains in which were deposited sands and muds from the elevated lands to the north and west. Swamps and local forests took the place of the epeiric sea. The terrestrial clastic sediments and lacustrine limestones laid down at that time now comprise the Kootenai formation. The Saypo area, however, was not much elevated, and the rocks were not deformed at the end of the Jurassic. Consequently, the Kootenai is so conformable on the Ellis that the contact between the two formations is difficult to locate.

The Upper Cretaceous was a time of alternating marine and terrestrial sedimentation in the southwestern part of the Saypo quadrangle and

also was the last time that marine waters entered the area. At present the youngest Upper Cretaceous rocks belong to the basal part of the Colorado shale. Sedimentation may have continued in the southwestern part of the Saypo area until near the close of the period (Fox Hills time), in which case the Montana group was removed by erosion after the Laramide orogeny. Magmas which cooled to form diorite sills were intruded into the Kootenai and Colorado sediments late in the Cretaceous or early in the Tertiary before the first orogenic movements began.

In summary, the region today occupied by the Lewis and Clark and Sawtooth ranges in the southwestern part of the Saypo quadrangle alternately received sediments or lay just above sea level from late in the Proterozoic until the end of the Mesozoic. During the times of elevation the area experienced very little erosion. The Laramide revolution, which produced most of the present geologic structures in the area, ended the extremely long period of quiet sedimentation.

SEQUENCE DURING LARAMIDE OROGENY

General statement.—Evidence is not available in the southwestern part of the Saypo quadrangle to determine accurately the age of the Laramide orogeny. Willis (1902, p. 343) postulated two periods of compression in northwestern Montana separated by "the Blackfoot (erosion) cycle" and dated the Lewis overthrust as mid-Tertiary. Mackenzie (1922, p. 115) suggested that the Laramide revolution "took place not earlier than the uppermost Cretaceous nor later than the latest Eocene", and that "the Lewis thrust was one of the latest effects of the compressive stresses of the Laramide revolution." Dake (1921, p. 245-254) summarized the evidence bearing on the problem in northwestern Wyoming and concluded that the Laramide orogeny developed in three episodes or stages which he assigned to the intervals between the Lance and Fort Union, between the Fort Union and the Wasatch, and after the Bridger. The evidence given by Dake indicates that most of the deformation took place after the close of the Cretaceous and before the close of the Eocene. Although the two diastrophic phases of the orogeny in the Saypo quadrangle are not correlative with those known for Wyoming, the fact is indicated that the orogeny also was multiple in the Lewis and Clark and Sawtooth ranges.

Early compressional phase.—The structures in the Saypo quadrangle indicate clearly that the deforming stresses acted from west to east. Because all the earlier structures were the result of compression and the later ones of tension, they may be divided into two phases. The

first event in the earlier or compressional phase was gentle folding of the 14,500 feet or more of sedimentary rocks in the Cordilleran and overlapping Rocky Mountain geosynclines. The flexures may have been nearly completed shortly after the Horsethief sandstone (Stebinger, 1918, p. 154) was deposited. The Kootenai and Colorado sediments and diorite sills probably were folded at this time. If the folding in Patrick's Basin was the earliest event of deformation, the sills must have been injected between the cessation of Kootenai sedimentation and long before the beginning of the orogeny because they were already cold when they were folded. After the rocks were first folded the region now occupied by the mountains undoubtedly was elevated more than the plains area to the east. Erosion stripped the younger and softer Upper Cretaceous shales and sandstones from the uplifted area. In this way part of the sediments above the Colorado shale may have been removed before faulting.

The second event in the older phase of the Laramide revolution in the Saypo quadrangle was high-angle thrust faulting which broke the rocks into a series of blocks or slices in the area now occupied by the Sawtooth Range. The slices consisted of Cambrian, Devonian, Mississippian, Jurassic, and Cretaceous sediments (Fig. 3, eastern part of sections). Each slice was tilted westward as it rode up over the contiguous eastern slice along a high-angle thrust which, like the strata, dipped west. Apparently the slices piled against each other like a deformed pack of cards and probably rode eastward on a sole (Cadell, 1888, p. 347-378) from which each high-angle thrust extended upward and increased in steepness. The thrusts in the eastern and western parts of the Sawtooth Range dip less steeply than those in the central part of the range. This fact seems to refute the suggestion that the slices were rotated above the soles (Mackenzie, 1922, p. 129) "so that the faults and the strata were given westward dips steeper than those caused by compression alone." The western slices should have been rotated through a larger angle than the eastern ones if rotation took place after the sole was formed, but the western faults dip less steeply than any of the others (Fig. 3, C-C′, central-western and eastern parts of sections).

The third event in the older phase of deformation was the development of low-angle overthrusts. As noted earlier in this paper, four such faults are present above the Lewis overthrust in the southwestern part of the Saypo quadrangle. The fact that the Lewis overthrust truncates the westernmost high-angle thrust indicates conclusively that the low-angle overthrusting was later than the high-angle thrusting. Whether the high-angle thrusts were developed farther west than they are ex-

posed, and lie beneath the Belt rocks west of the trace of the Lewis overthrust, is unknown. The length of time between the high- and low-angle faulting also is unknown. When the last overthrusts were completed the mountains probably stood at their greatest heights and the maximum amount of shortening within the central part of the Cordilleran geosyncline was completed. The folding and faulting of the Mesozoic rocks in the plains area adjacent to the mountain front (Stebinger, 1918, p. 168, 171-172) may have been completed when the low-angle overthrusts were formed west of the Sawtooth Range, the stress being transmitted eastward through the then-compressed mass of slices.

Late tensional phase.—The second phase of the Laramide orogeny within the Saypo quadrangle was adjustment of the disturbed strata after the compressive forces from the west relaxed. The length of time between the two phases is unknown. The first event during the second phase, and the fourth in the Laramide revolution, was the development of an eastward-dipping low-angle normal fault which dropped Middle Cambrian limestones against the sandstone of the Hoadley between Glenn and Moose creeks (Fig. 3, A-A′, west part of section).

The fifth and final event was the development of a series of high-angle westward-dipping normal faults which, in the western part of the quadrangle, broke the low-angle normal fault and also the underlying overthrusts. These high-angle faults probably have an average stratigraphic displacement of 2000 feet, indicating much settling and readjustment of the rocks after the compressive stresses ceased. The adjustments appear to have been completed early in the Tertiary. Evidence was not found which suggests that any significant structures were formed later.

POST-LARAMIDE SEQUENCE

Erosion has been continuous since the Laramide revolution in the Sawtooth and Lewis and Clark ranges. During the Tertiary most of the Upper Cretaceous rocks were eroded from the mountains, and streams carried the detritus eastward across the plains. In the Lewis and Clark Range the Mesozoic deposits were removed completely, and many of the Paleozoic rocks also were eroded before the Pleistocene.

The mountains must have been greatly lowered by the end of the Miocene, or before the final elevation of the northern Rockies during the Pliocene. The slight variation in altitudes of the existing peaks suggests that the mountains were partly peneplained before the uplift late in the Tertiary. It is difficult to visualize how such streams as

Deep Creek and the Teton and Sun rivers could have attained their east-west course across the strike of the mountain ridges unless the region were eroded nearly to base level. The final uplift of the northern Rockies in Montana apparently did not produce any significant folding or faulting.

Immediately following the uplift, erosion cut down the soft shales and sandstones of the Kootenai and Ellis more rapidly than the hard limestones of the Cambrian, Devonian, and Mississippian and produced the long straight valleys and the dividing ridges which characterize the Sawtooth Range. Thus the present drainage pattern was established before the Pleistocene, when snow and ice collected in the valleys to form the first glaciers. The exact time in the Pleistocene when the first alpine glaciation occurred is unknown, but the ice filled the valleys and covered the low ridges. Erratics from the Lewis and Clark Range were carried east and dropped on the ridges below an altitude of 7000 feet (Pl. 2, fig. 2). The mountains must have been covered with ice so that only the higher peaks projected above the valley glaciers. The glaciers carved cirques in the higher peaks and deepened the valleys and cut them into U-shaped profiles. When the glaciers retreated, they left lateral and recessional moraines at various places in the mountains.

The second glacial stage (late? Wisconsin) recarved the peaks and valleys, and locally removed part of the older moraines. When the Wisconsin ice melted, fluvio-glacial sands and gravels were deposited in many of the valleys. Concerning the glacier which flowed east from the Sawtooth Range through the Sun River Canyon, Alden (1932, p. 120) said:

> "The piedmont glacier on the North Fork of the Sun River in Lewis and Clark and Teton Counties was exceeded in size only by the great Two Medicine Glacier. As shown by the moraines mapped by Stebinger, the ice filled the "gateway" or "portal" (pl. 43) and overrode the top of the lower mountain wall (pl. 7 A) as a valley glacier 4 miles wide. Passing the foothills it deployed on the gently undulating plain until it reached a maximum width of about 15 miles, north to south, and a length of 18 miles."

Since the retreat of the Wisconsin ice the streams in the mountains have cut narrow sheer-walled canyons in the bedrock of the U-shaped valleys in some places and deposited alluvium in others where the velocity locally is reduced.

CONCLUSIONS

The oldest rocks in the quadrangle are exposed only in the Lewis and Clark Range, are 8200 feet thick, are late Belt (pre-Cambrian) in age, and can be divided into four formations which, from oldest to youngest, are the Miller Peak argillite, Cayuse limestone, Hoadley formation,

and Ahorn quartzite. The Belt formations are correlated with the lower half or two-thirds of the Missoula group (Clapp and Deiss, 1931, fig. 2) in the Sapphire and Garnet ranges. The Cayuse, Hoadley, and Ahorn formations are named and described for the first time.

The Cambrian rocks in the region are Middle and Upper Cambrian and have an average thickness of 1800 feet, the lower 350 feet being sandstone and shale and the upper 1450 feet being limestone, dolomite, and a little shale. The Cambrian rocks are all assigned to the nine formations previously described from northwestern Montana (Deiss, 1939, p. 34-47).

The Devonian system is represented by 980 feet of Upper Devonian limestone, dolomite, and shale (*see* Allan Mountain section) which probably is not precisely correlative with the Jefferson and Three Forks formations in central Montana. The Devonian rocks have been mapped as two formations over much of the quadrangle but the rocks are not equivalent to the Jefferson and Three Forks formations in adjacent regions. The Devonian rocks in the Saypo area must be given new formational names when the geologic map is published.

The Carboniferous system is represented by Mississippian limestone, dolomite, and shale, estimated to be 1370 feet thick. The rocks probably are equivalent to both the Madison and Brazer formations of Montana and Utah. Because of this fact and because the Mississippian rocks must be mapped as a unit they are described and named as a new formation, the Hannan limestone.

The Jurassic system is represented by 250 feet or more of shale, limestone, and sandstone which are tentatively referred to the Ellis formation, although the rocks probably are equivalent in age to only the upper part of the Ellis in central Montana. A new stratigraphic section of these rocks was measured at the head of Lime Gulch. The fossils collected from the section were identified by Reeside who assigned them to the Divesian and Argovian.

The Lower Cretaceous rocks all belong to the Kootenai formation which, within the map area, consists of alternating red and green shales and drab-colored sandstones. The upper-middle part of the formation contains one or more lenticular beds of chert-pebble conglomerate and lacustrine fossiliferous limestone. The formation apparently is terrestrial.

The Upper Cretaceous is represented in the Sawtooth Range by the lower part of the Colorado shale which is the youngest indurated formation recognized in the area. Most of the rocks in the formation are correlated with the Blackleaf sandy member of the Colorado east of the mountains. The brachiopods and pelecypods, collected from the upper part of the Blackleaf in the Sawtooth Range, indicate correlation

of the rocks with part of the Aspen and Bear River formations of Wyoming.

Angular unconformities are not recorded in the Saypo quadrangle, but one small and three large hiatuses are known. The greatest angularity occurs between the Cambrian and Belt formations where the discordance of dip may be as much as 5°. The other two large hiatuses are between the Upper Devonian and Middle Cambrian, and between the Upper Jurassic and Mississippian formations. Other gaps in the geologic column probably are present but have not been observed.

Faults are the principal geologic structures in the area, but some folds also are present. The structures were formed during the Laramide revolution, either during the latest Cretaceous or early Tertiary.

Open folds were formed, possibly as the first event in the Laramide orogeny, and small overturned folds were dragged against some of the thrust faults. The tension fractures in the crests of the open folds in the diorite sills in the Kootenai clearly indicate that the magmas were intruded and cooled before deformation.

Four types of faults are recognized in the area and appear to have originated in the order named: (1) westward-dipping high-angle dip-slip strike thrusts which produced blocks or slices of Paleozoic and Mesozoic rocks that moved eastward on a sole; (2) westward-dipping low-angle dip-slip strike or slightly oblique overthrusts, including the Lewis overthrust, which broke Belt and older Paleozoic rocks; (3) one eastward-dipping low-angle dip-slip oblique normal fault which dropped Cambrian limestones on Hoadley (Belt) sandstone; (4) westward-dipping high-angle dip-slip oblique and strike normal faults which dropped Cambrian against Belt, Devonian against Cambrian, and Mississippian against Devonian.

Erosion during the Tertiary may have peneplained the mountains. The area was re-elevated in the Pliocene? but without pronounced folding or faulting.

Pleistocene alpine glaciers and Recent streams have produced most of the existing topographic details and are largely responsible for the ruggedness of the mountains.

REFERENCES CITED

Alden, W. C. (1932) *Physiography and glacial geology of eastern Montana and adjacent areas,* U. S. Geol. Survey, Prof. Paper **174**, 133 pages, 51 plates, 19 figures.

Bell, W. C. (1941) *Cambrian Brachiopoda from Montana,* Jour. Paleont., vol. 15, p. 193-255, pls. 28-37, figs. 1-20.

Bevan, Arthur (1929) *Rocky Mountain front in Montana,* Geol. Soc. Am., Bull., vol. 40, p. 427-456, 8 figures.

Blackstone, D. L., Jr. (1934) *Brachiopoda from the Madison limestone in Montana* (Unpublished M. A. Thesis), Mont. State Univ., 110 pages, 7 plates.

Cadell, H. M. (1888) *Experimental researches in mountain building,* Royal Soc. Edinburgh, Tr., vol. 35, p. 337-357, il.

Chapman, R. H. (1900) *Notes on the structure of the Rocky Mountains in the Lewis and Clarke Timber Reserve, Montana,* Am. Inst. Min. Engs., Tr., vol. 29, p. 153-156, 2 figures (1899).

Clapp, C. H. (1932) *Geology of a portion of the Rocky Mountains of northwestern Montana,* Mont. Bur. Mines Geol., Mem. no. 4, 30 pages, map.

—— (1934) *Structure of the Coopers Lake quadrangle, Montana* (abstract) Geol. Soc. Am., Pr., p. 72.

——, and Deiss, C. F. (1931) *Correlation of Montana Algonkian formations,* Geol. Soc. Am., Bull., vol. 42, p. 673-696, 3 figures.

Clapp, M. M. (1936) *Beltian stratigraphy and structure in southern part of Ovando quadrangle, Montana* (Unpublished M. A. Thesis), Mont. State Univ., 45 pages, 3 plates, 1 figure.

Collier, A. J. (1927) *The unconformity between the Madison limestone and Ellis formation in northern Montana* (abstract), Washington Acad. Sci., Jour., vol. 17, no. 5, p. 130-131.

Crickmay, C. H. (1936) *Study in the Jurassic of Wyoming,* Geol. Soc. Am., Bull., vol. 47, p. 541-564.

Crosby, W. O., *in* Powers, Sidney, and Shimer, H. W. (1914) *Notes on the geology of the Sun River district, Montana,* Jour. Geol., vol. 22, p. 556-559, 1 figure.

Dake, C. L. (1921) *Episodes in Rocky Mountain orogeny,* Am. Jour. Sci., 5th ser., vol. 1, p. 245-254.

Daly, R. A. (1912) *Geology of the North American Cordillera at the forty-ninth parallel,* Geol. Survey Canada, Mem. no. 38, pt. 1, p. 1-546, pls. 1-73, figs. 1-42.

Dawson, J. W. (1885) *On the Mesozoic floras of the Rocky Mountain region of Canada,* Science, vol. 5, p. 531-532.

—— (1886) *On the Mesozoic floras of the Rocky Mountain region of Canada,* Royal Soc. Canada, Pr., Tr. (1885), sec. 4, p. 1-22, 1 figure.

Deiss, Charles (1933) *Paleozoic formations of northwestern Montana,* Mont. Bur Mines Geol., Mem. no. 6, 51 pages, 3 plates (Mimeographed).

—— (1935) *Cambrian-Algonkian unconformity in western Montana,* Geol. Soc Am., Bull., vol. 46, p. 95-124, pls. 7-8.

—— (1939) *Cambrian stratigraphy and trilobites of northwestern Montana,* Geol Soc. Am., Spec. Paper 18, 135 pages, 18 plates, 7 figures.

—— (1940) *Lower and Middle Cambrian stratigraphy of southwestern Alberta and southeastern British Columbia,* Geol. Soc. Am., Bull., vol. 51, p. 731-794, 5 plates, 6 figures.

—— (1941) *Cambrian geography and sedimentation in the central Cordilleran region,* Geol. Soc. Am., Bull., vol. 52, p. 1085-1116, 10 figures.

Girty, G. H., *in* Mansfield, G. R. (1927) *Geography, geology, and mineral resources of part of southeastern Idaho,* U. S. Geol. Survey, Prof. Paper 152, p. 63-71.

Knowlton, F. H. (1907) *Description of a collection of Kootenie plants from the Great Falls coal field of Montana,* Smithson. Misc. Col., vol. 50, Quart. Issue 1, p. 105-130, pls. 11-14.

Mackenzie, J. D. (1922) *The historical and structural geology of the southernmost Rocky Mountains of Canada,* Royal Soc. Canada, Tr., 3d ser., vol. 16, sec. 4, p. 97-132, 3 figures.

Mansfield, G. R. (1927) *Geography, geology, and mineral resources of part of southeastern Idaho,* U. S. Geol. Survey, Prof. Paper 152, 453 pages, 70 plates, 46 figures.

Peale, A. C. (1893) *The Paleozoic section in the vicinity of Three Forks, Montana,* U. S. Geol. Survey, Bull., no. 110, 56 pages, 6 plates.

Powers, Sidney, and Shimer, H. W. (1914) *Notes on the geology of the Sun River district, Montana,* Jour. Geol., vol. 22, p. 556-559, 1 figure.

Reeside, J. B., Jr. (1919) *Some American Jurassic ammonites of the genera* Quenstedticeras, Cardioceras, *and* Amoeboceras *family* Cardioceratidae, U. S. Geol. Survey, Prof. Paper 118, 64 pages, 24 plates.

Richardson, G. B. (1913) *The Paleozoic section in northern Utah,* Am. Jour. Sci., 4th ser., vol. 36, p. 406-416.

Sanderson, J. O. G. (1931) *An Ellis (Upper Jurassic) section at East Butte, Sweetgrass Hills, Montana,* Am. Assoc. Petrol. Geol., Bull., vol. 15, p. 1157-1160, 6 figures.

Stebinger, Eugene (1918) *Oil and gas geology of the Birch Creek—Sun River area, northwestern Montana,* U. S. Geol. Survey, Bull., no. 691-E, p. 149-184, pl. 24, figs. 32-34.

Willis, Bailey (1902) *Stratigraphy and structure, Lewis and Livingston ranges, Montana,* Geol. Soc. Am., Bull., vol. 13, p. 305-352, pls. 48-53, 5 figures.

Montana State University, Missoula, Montana.
Manuscript received by the Secretary of the Society, January 31, 1942.
Presented before Cordilleran Section of the Geological Society, April 19, 1941.
Published by permission of the Director, U. S. Geological Survey.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 54, PP. 263-286, 4 FIGS. FEBRUARY 1, 1943

ELASTICITY OF IGNEOUS ROCKS AT HIGH TEMPERATURES AND PRESSURES

BY FRANCIS BIRCH

CONTENTS

	Page
Abstract	263
Introduction	263
Acknowledgments	265
Method	265
Description of materials	270
Results	273
References cited	284

ILLUSTRATIONS

Figure	Page
1. Original readings for bronzitite at 8000 kg/cm²	273
2. Original readings for New Zealand dunite at 8500 kg/cm²	276
3. Relative frequency as function of temperature for various rocks at high pressure	278
4. Relative frequency as function of temperature for various rocks at high pressure	280

ABSTRACT

In a continuation of earlier studies of the elasticity of rocks at high pressures, the rigidity and the velocity V_s of shear waves have been determined for a number of igneous rocks at temperatures up to 600° C, at pressures of 3000 to 9000 kg/cm². The extreme experimental conditions approach closely the conditions of pressure and temperature to be found at the base of a 35-kilometer crust. The new measurements are presented briefly; a thorough discussion of their significance is deferred for a more complete future publication.

INTRODUCTION

For interpreting the structures discovered by seismological research, which are described in terms of velocities of various elastic waves, we require a systematic body of data from which the velocities for any known or postulated type of rock may be obtained with a precision at least comparable with that of seismological measurements. In particular, in order to deal with the several layers constituting the uppermost 30 to 50 km, the "crust", we require studies of the elasticity of rocks under conditions approaching those which exist at such depths. One fundamental char-

acteristic of these conditions is the existence of a mean compressive stress increasing with depth and reaching about 10,000 atmospheres at 35 km. As a consequence, the most relevant measurements for use in efforts to identify the constituents of the crustal layers have been the various measurements at high pressure (Adams and Williamson, 1923; Adams and Gibson, 1926; 1929; Bridgman, 1924; 1925; 1928; Birch and Bancroft, 1938; 1940).

An element of no less importance is temperature, which increases with depth at an average rate still in question, but, at least near the surface, of an order sufficiently well known to afford a rough estimate of the possible temperatures in the crust. It seems likely that at a depth of 35 km the mean temperature is between 500° and 800° C. The effect of such temperatures upon the elasticity of rock-forming minerals (except quartz) and of rocks under high pressure has not hitherto been investigated, and ignorance of this effect has introduced an unavoidable and, as we shall see, a very appreciable element of error in our attempts to utilize a juxtaposition of seismic velocities with laboratory velocities for identifying the crustal layers. The measurements described below were intended to eliminate, insofar as possible, the necessity for extrapolation in dealing with the upper 35 km (22 miles) of crust.

In principle, the new measurements depend upon methods already used in the determination of the velocity of waves of shear, or of the rigidity in rocks at high pressures and at temperatures up to 100° or 150°C. (Birch and Bancroft, 1938; 1940). A few changes of technique have resulted in the extension of the useful range of temperature to nearly 700°C. The measurements of velocity are obtained as function of temperature at various constant pressures which may be as high as 10,000 kg/cm^2. Earlier studies have shown very definitely that no satisfactory measurements of elasticity can be obtained on heating rocks at ordinary pressure (Ide, 1937); at temperatures above 100° C. or even lower, rocks begin to disintegrate as a result of the thermal expansion of anisotropic crystals. This disintegration can be very largely prevented by the application of high pressures over the exterior surface of the rock. As we shall see, measurements at high pressures give for rocks effects of temperature which are probably of the same order as those for the individual crystals.

These studies have not yet been carried out on the scale originally planned, but it seems desirable to publish the partial results with the hope of adding a greater variety of materials and conditions when circumstances permit a resumption of the work.

ACKNOWLEDGMENTS

This work was financed by a grant from the Penrose Bequest of The Geological Society of America. The writer is indebted to Professor E. S. Larsen for the specimens of eclogite and diorite and for a number of the analyses; to Mr. George Switzer for analyses of the eclogite; to Mr. J. C. Rabbitt for collecting and analysing the amphibolite; and to Mr. George C. Kennedy for an analysis of the quartz diorite. A fine specimen of the Dun Mountain dunite was kindly shipped from New Zealand by Dr. W. N. Benson, of the University of Otago, Dunedin. The help of Professor F. B. Hunt in selecting a permanent magnet for use at high temperatures is gratefully acknowledged.

METHOD

The effective rigidity G is related to the velocity of propagation of waves of shear V_S according to the formula, $V^2_S = G/\rho$. The lowest resonant frequency of torsional vibration of a uniform circular cylinder is given by $f = V_S/2\,L$, where L is the length. Thus the determination of this frequency for cylinders of known length and density gives both V_S and G. It is not necessary that the material be isotropic for the application of these expressions, but it must be uniform in the direction of the axis of the cylinder. Actually, our materials are uniform in the sense that one section of sufficient length is approximately like another of the same length; the cylinders cannot be uniform on a microscopic scale. The same remark applies to the application to seismology of any of the results derived from the theory of the elasticity of a homogeneous continuum.

The vibrations of the cylinders are produced and observed electromagnetically. A small bar magnet is fixed across each end of the cylinder. The ends of one magnet are in air gaps in which there is a sinusoidally varying magnetic field of controllable frequency; a torque of the frequency of the field is exerted on the magnet, and thence upon the end of the cylinder. The motion of the magnet at the other end induces a voltage of the same frequency in another set of coils; these coils are connected to amplifiers and a voltmeter, and resonance is found by observing the variation of the induced voltage as the driving frequency is varied. This is the same method, in principle, which has been described in previous publications.

In the earlier arrangement, the bar magnet was a piece of soft iron, polarized by a permanent magnet on which the coils were mounted. It was now found possible to reduce the driving and pickup units very greatly in size by putting the permanent magnet on the specimen. At

the same time, the number of coils at each end was reduced from 4 to 2. The result was a driving unit less than ¾ inch in diameter and 2 inches long. It is essential to use a magnetic material which retains an appreciable fraction of its strength at the highest temperature to which the specimen is to be exposed; this means a material with high coercive force, and, if possible, a high remanent magnetism. At the suggestion of Professor F. B. Hunt, a material known as "Nipermag" was tried and found to be surprisingly satisfactory. The Curie temperature for iron is about 760°C.; the "Nipermag" permanent magnets were still sufficiently magnetic at 660°C. to drive specimens of rock.

The coils were of No. 20 asbestos-insulated copper wire, wound on pole pieces made from transformer iron sheets. Each coil consisted of two layers, with about 21 turns in each layer. The low impedance of these coils was matched to the driving amplifier and to the pickup amplifier with suitable matching transformers.

The diameter of the specimens being uniformly ⅝ inch, it was possible to slip the specimen, with the driving and pickup coils, into a brass tube of ¾-inch internal diameter; this left $\frac{1}{16}$-inch radial clearance about the specimen, which was mounted coaxially in the tube upon coaxial steel pins soldered to the centers of the magnets and held in holes drilled along the axis of the driving and pickup units. The outside diameter of the brass tube was 0.810 inch; this tube was made to slide easily into another thin-walled brass tube on which, insulated with sheet mica, was wound a heating element of nichrome ribbon. The furnace tube in turn was wrapped with asbestos paper to a diameter of nearly 1.5 inches and contained in a third thin-walled brass tube, having an outside diameter of 1.5 inches. This whole assembly was placed in a thick-walled steel pressure vessel.

Electrical connections to the driving and pickup units, to the heating coil, and to the thermoelements used for measuring the temperature were made through insulated pressure-tight leads of a type already described. Pressure was built up with a 200-ton hydraulic press (Birch and Law, 1935). The pressure medium was nitrogen; this was taken from commerical cylinders at a pressure of from 1000 to 2000 lbs/in², compressed to 15,000 lbs/in² in a preliminary stage, and finally to pressures of as much as 10,000 atmospheres by use of the hydraulic press. Pressure was measured by the change of resistance of a manganin gauge coil placed in a recess which was not heated by the furnace.

The resistance of the furnace winding was about 10 ohms. About 1000 watts were required for heating to 600°C. with high pressures in the cylinder, a considerably greater power consumption than was required with 1 atmosphere of pressure. The compressed nitrogen is a rather

good medium for the transfer of heat. The cylinder and specimen were mounted horizontally, and the tight layers of asbestos paper diminished convection. The steel cylinder gradually warmed up to a temperature at which it could dissipate at its surface the power supplied to the furnace. This temperature proved to be a moderate one, and the only cooling device employed was the circulation of air about the cylinder by small fans.

The distribution of temperature within the furnace was investigated by means of two thermocouples, one with its hot junction near the center of the specimen, the other with its hot junction near the end of the specimen. The difference in temperature between these two points increased with the mean temperature but did not exceed 40° when the temperature at the center was 600°; the temperature difference within the specimen itself was probably somewhat less. This difference is not enough to have a serious effect upon the results. The frequency of vibration is principally dependent upon the elasticity near the center of the specimen, the parts near the ends behaving almost as if they were masses without elasticity. If we take a case where the effect of 600° is to change the frequency by 10 per cent (about the maximum observed effect), then 40° would correspond to 10/15 or 0.6 per cent of the frequency if the whole specimen changed temperature by this amount; where this is the difference between the ends and the middle, the resulting change of frequency is much less. A calculation for a sinusoidal distribution of temperature with 40° as the greatest variation gives a change of frequency of only 0.15 per cent.

The temperature was always measured at the center of the specimen and usually at one other point. The thermocouple leads passed through insulated packings (Birch, 1939) to the cold junction, from which copper connections were carried to a Type K potentiometer. Chromel-alumel thermocouples were used, in conjunction with the manufacturer's calibration table; the absolute precision of the temperature measurement was thus about 1 per cent, which was ample for this work. For some of the measurements, a Wheelco controller was used to control the furnace temperature; the mercury switch of this controller was shunted by a suitable resistor so that the controller turned off and on a small fraction of the total heating current. In this way, violent fluctuations were eliminated and satisfactory control established. It was generally possible, however, to eliminate control and to allow the furnace temperature to follow the line voltage; except at certain hours of the day, the changes of temperature with a given setting of the Variac which determined the heating voltage were small and slow enough to cause no difficulty.

The thermal inertia of this furnace was very small, since there was no large mass of insulating material; it was possible to heat very rapidly, and, due to the action of the dense nitrogen, equilibrium was rapidly attained. It was necessary to allow a little time for the specimen itself to reach the temperature of the furnace, but for $\frac{5}{8}$-inch specimens this time was about 5 or 10 minutes. Measurements could thus be made very rapidly, except that it was desirable to prolong the time of heating at each temperature to permit as much readjustment of the crystals as possible. A change of temperature, at high pressure as well as at low, tends to loosen the structure of an aggregate of noncubic crystals; in time, at sufficiently high pressure, the processes of flow tend to restore a compact structure. These processes are accelerated at high temperatures, but it is desirable to prolong the heating to the practical limit. Unfortunately, this has been a matter of a day or so, but it is not believed that longer time would have had a significant effect in most cases.

The pressure of the nitrogen must be exerted only upon the external surface of the rock cylinder; it must not be allowed to penetrate the intercrystalline spaces. This restriction can be accomplished by enclosing the rock cylinder in a sheath of copper foil. For the present work, it was, of course, essential that this foil be pressure-tight at the highest temperatures, and, consequently, the formerly satisfactory method of soldering the jacket with soft solder had to be abandoned. Very fortunately, hydraulically formed seamless copper tubes of the right size and wall thickness were available at the Clifford Manufacturing Company. One end of these tubes was already closed, and it was thus necessary only to spin a copper cap for one end and to solder this and the magnets with hard solder. Without these or similar tubes, it would have been very difficult indeed to accomplish these measurements. Even with them, the closure of one end was not always successfully performed; in a rather large number of cases, leaks developed at the soldered joint, and in a few cases punctures were caused by the pressure along the unsoldered portion of the tubes. When once satisfactorily achieved, however, these jackets were remarkably durable and permitted, in a number of cases, several repetitions of measurements to 600° or more.

The jacketed specimen, magnet, and mounting pin were aligned in a suitable jig and hard soldered together. Since all of the soldering was at the ends of the specimen, relatively little damage was done the specimen by the rather severe heating at 1 atmosphere; the important central portion was not allowed to rise to an uncomfortably high temperature. It is probable, however, that this heating was partly responsible for the lower values of velocity now found for some of the specimens which had been previously studied without this mistreatment.

The jackets now used were nearly twice as thick as those used in the former experiments, and the magnets were about twice as heavy. The corrections necessary for getting the absolute frequency of the specimen alone were consequently somewhat higher than formerly—about 8 or 9 per cent, as against 5 to 6 per cent before. These corrections were found by calculation by Rayleigh's method, and also by measurement. The uncertainty of the absolute velocity does not exceed 2 per cent and is usually no greater than 1 per cent. The uncertainty of the relative frequencies is less than this, since the corrections remain very nearly constant as the temperature changes. The relative frequencies are thus dependent only upon the precision with which the relative capacities in the oscillator corresponding to the resonant frequencies can be determined. This is of the order of 1 part in 1000 and is ample in view of the uncertainty and lack of uniformity of the temperature. The mean temperature coefficients are known to within 1 to 5 per cent, insofar as the precision of the measurements alone is concerned. On the other hand, the changes which may take place in the rocks themselves introduce uncertainties which may be considerably greater than this. Some idea of the magnitude of these effects is given in later sections.

The measurements were generally conducted in the following way. The pressure was first raised, at room temperature, by steps of about 1000 kg/cm^2; readings of resonant frequency were taken at each step. When the desired pressure was reached, it was maintained constant while the temperature was raised, by steps of from 60° to 100°C., to the desired maximum, or to a temperature at which some accident halted the run—a leak in the jacket being by far the most common accident. Normally readings also were made on cooling, by similar steps, to room temperature. The pressure was adjusted as necessary to a constant value for the whole run—that is, to a value within about 200 kg/cm^2 of the nominal value. Frequently it happened that at some temperature on the first heating a reading was obtained which was evidently out of line with those at lower temperatures; the procedure was then to return to lower temperatures to find whether this was a reversible effect, and then to raise the temperature again beyond this point. Wherever possible, the whole curve was retraced, from room temperature to the maximum. In several cases, this whole procedure was carried out at two pressures for a given rock. There is little evidence as yet that the pressure, when greater than a few thousand atmospheres, has much effect upon the temperature coefficient. In most cases, measurements were continued until a failure of some kind took place. In order to have some idea of the effect of duration, the pressure was generally maintained over night, with a moderate temperature, perhaps 200° to 300°C. This seemed to

have little effect. Such changes as took place generally happened quickly, and prolonged waiting, within the rather short possible intervals, had little effect.

The oscillator used for this work was the General Radio Low Frequency Oscillator 377-B used for former measurements. As the duration of these runs might be several days, the calibration of this oscillator was checked at frequent intervals against the Cruft Laboratory Standard 1000-cycle frequency. The drift of the oscillator frequency, for which a correction was made, was never greater than about one part in 1000.

DESCRIPTION OF MATERIALS

Most of the rocks used in these investigations have already been described (Birch and Bancroft, 1938; 1940; Birch and Clark, 1940). A correction to the published analysis of the Stillwater anorthosite has kindly been furnished by Professor H. H. Hess; the plagioclase is bytownite ($Ab_{15}An_{85}$) instead of $Ab_{35}An_{65}$, as originally given. Descriptions of the new rocks are given below.

AMPHIBOLITE, from South Boulder Creek, Madison County, Montana. This sample was part of a large boulder, collected with the assistance of Mr. J. C. Rabbitt, who has given the following description: a holocrystalline, coarse-grained recrystallization texture, consisting of irregular and subhedral grains of monoclinic amphibole (tremolite-actinolite) with interstitial orthorhombic amphibole (anthophyllite), dark-green spinel, yellow and gray serpentine, magnetite, a few grains of garnet. Monoclinic amphibole in irregular grains ranging up to 2 mm in greatest dimension; average grain size, 1.5 mm, with little or no elongation; pleochroic from colorless to faint green; $\gamma = 1.650$; $(-)2V$ large; $z \wedge c\ 15°$. The crystals contain abundant needlelike inclusions of magnetite (ilmenite) in a grating pattern with needles oriented in two directions, parallel to x and z, and few small grains of dark-green spinel. Most grains are fresh; some show incipient alteration to serpentine. Orthorhombic amphibole appears as irreguar grains ranging up to 0.75 mm in length and averaging 0.5 mm in diameter. This mineral, also forming irregular aggregates bounded by the monoclinic amphibole, is pleochroic from colorless to faint pink; $\beta = 1.655$; $(-)2V$ is large. Extensively altered to serpentine. Rosiwal analysis gives for the proportions by volume in percentages:

Monoclinic amphibole	75
Orthorhombic amphibole	11
Serpentine	6
Green spinel, Magnetite, Ilmenite (?)	8

DUNITE, from the type locality, Mt. Dun, New Zealand, is nearly pure olivine, with the composition 91 per cent of Mg_2SiO_4, 9 per cent of Fe_2SiO_4. An examination by Larsen showed 3 per cent of serpentine,

TABLE 1.—*List of materials*

ROCK	DENSITY
Quartzite, Georgia	2.647
Albitite, Pennsylvania	2.62
Anorthosite, Stillwater Complex, Montana	2.74
Bronzitite, Stillwater Complex, Montana	3.27
Amphibolite, Montana	3.14
Dunite, New Zealand	3.255, 3.237, 3.236, 3.262
Dunite, North Carolina	3.269, 3.252, 3.255, 3.275
Marble, Vermont	2.71
Granite, Barre, Vermont	2.65
Granite, W. Chelmsford, Massachusetts	2.62
Granite, Rockport, Massachusetts	2.62
Quartz diorite, California	2.81
Diabase, Maryland	2.96
Gabbro, Wisconsin	2.90
"Eclogite" (hornblende-garnet) California	3.44
Eclogite (omphacite-garnet) California	3.28, 3.31

1 per cent of chromite. A number of dark, thin seams traverse this material, which is, however, mechanically strong and much less friable than the coarser-grained, sugary dunite from North Carolina.

QUARTZITE DIORITE, from the San Luis Rey quadrangle, California, was collected by Larsen. This is fresh and coarse-grained, consisting of plagioclase blocks, with irregularly shaped grains of hornblende and biotite with interstitial quartz. The grain size of the principal constituents varies from 4 mm., to 0.2 mm., averages about 2 mm. The feldspar shows a slight development of sericite. The proportions (in percentages) by volume are as follows:

Quartz	19
Andesine ($Ab_{50}An_{50}$)	50
Biotite	17
Hornblende	13
Sericite, Magnetite, Apatite, Sphene, Zircon	1

"ECLOGITE" (hornblende-garnet granulite). This was collected by Larsen at the W.P.A. quarry near Occidental, Sonoma County, California. From examination of two thin sections, Mr. George Switzer gives the following description: The rock is granoblastic, with grain size between 0.06 and 2.0 mm. Hornblende and garnet are the chief constituents;

clinozoisite, muscovite, glaucophane, sphene, and chlorite are present in minor amounts. The composition of the garnet is given in Table 2, taken from Pabst (1938). The proportions (in percentages) by volume are:

Hornblende	63
Garnet	20
Muscovite	4
Clinozoisite	3
Sphene	5
Chlorite	4
Glaucophane	1

ECLOGITE (omphacite-garnet granulite). This was also collected by Larsen, from the W.P.A. quarry near Healdsburg, Sonoma County, California. The grain size ranges from 0.05 to 2.0 mm. Analysis by Larsen gave the following proportions (in percentages):

Pyroxene (or phacite)	72
Garnet	24
Sphene	2
Chlorite	1
Glaucophane	1

A chemical analysis of the pyroxene by F. A. Gonyer is shown in Table 2; it falls between jadeite and augite. The garnet is like the garnet of the other "eclogite". This rock is probably correctly designated as eclogite, although the amount of garnet is perhaps somewhat low.

TABLE 2.—*Chemical analyses*

	A	B	C	D
SiO_2	53.31	38.26	37.40	38.36
TiO_2	.26	.....	tr.	.....
Al_2O_3	10.52	19.93	1.54	.58
Fe_2O_3	4.11	4.87	1.89	1.58
FeO	2.84	20.40	6.36	6.62
MnO	.05	.....	.13	.09
MgO	8.42	3.94	49.92	51.50
CaO	14.50	12.02	tr.	.10
Na_2O	5.90	.....	tr.	.....
K_2O	.05	.....	tr.	.....
H_2O-	.....	.....	1.42	.....
H_2O+	.16	.....	.....	.....
Cr_2O_3	.....	.....	1.00	.....
NiO	.....	.....	.....	.34

A. Pyroxene from eclogite, F. A. Gonyer, unpublished analysis.
B. Garnet, Sonoma County, Calif., A. Pabst (1938).
C. Dunite, Mt. Dun, New Zealand, given by Dr. W. N. Benson.
D. Olivine in dunite from Balsam Gap, N. C.; F. A. Gonyer in Ross, Shannon, and Gonyer (1928, Analysis No. 2, p. 545).

QUARTZITE, Georgia. This is a very pure white medium-grained quartzite, obtained from the Quincy Granite Company. It contains more than 99 per cent of quartz.

Chelmsford granite. This granite, which is described by Dale (1923) was obtained at the H. E. Fletcher quarry, West Chelmsford, Mass. A new analysis by Larsen is given below (volume percentages):

Quartz	31
Oligoclase	31
Microcline, microperthite	31
Muscovite	4
Biotite	1
Chlorite	2

RESULTS

The numerical results are displayed in a number of figures and tables. A few examples of the original data are given, but for conciseness most

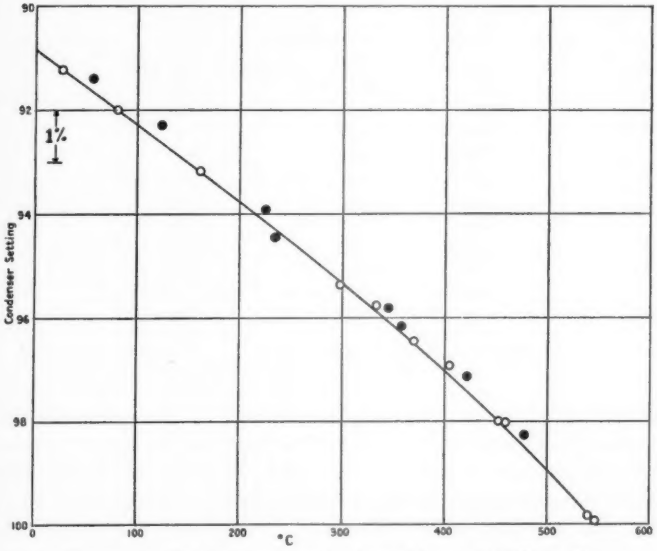


FIGURE 1.—*Original readings for bronzitite at 8000 kg/cm²*

(Condenser settings for resonance as function of temperature.) Open circles denote observations with increasing temperatures; solid circles, observations with decreasing temperatures.

of the tabulated data are smoothed values, given at 100-degree intervals of temperature, at various constant pressures.

One example of the original data is given in Figure 1 and Table 3. In the table, the readings of oscillator condenser setting at resonance, versus temperature, are given in the original order for the bronzitite; another example, the New Zealand dunite, is given in Figure 2 and Table 4. The bronzitite is a good example of a rock which behaves in the

TABLE 3.—*Original data for bronzitite at 8000 kg/cm²*

(Condenser setting versus temperature; the readings are given in the order in which they were taken.)

t°C	Capacity	Time
27	0.09123	July 9, 4:40 PM
81	9200	5:07
162	9317	5:30
234	9426	7:30
298	9532	8:00
370	9645	8:30
453	9800	8:50
358	9616	9:20
	overnight at 235°	
235	9425	July 10, 7:45 AM
334	9575	8:10
405	9692	8:45
460	9802	9:25
540	9982	9:45
547	9991	10:03
478	9826	10:55
422	9712	11:30
345	9581	12:10
225	9391	1:00 PM
124	9229	2:00
57	9139	2:50

Calibration: 0.09191 = 12,500 cycles/sec.
.09998 = 12,000 cycles/sec.

"normal" way: that is, a gradual improvement of the mechanical condition takes place so that, at the end of the run, the resonant frequency is slightly higher than it was initially. This behavior may reasonably be attributed to a gradual increase of compactness, a "squeezing out" of the original porosity, which in this rock is very small. The irreversibility shown here is so small as to have little effect upon the average temperature coefficient.

The dunite shows another kind of irreversible effect. At about 300°, the frequency of this rock dropped off about 1 per cent below a continuation of the line (broken line in Fig. 2) given by the readings at lower temperature; on cooling from this point, a new line (the solid line of

Fig. 2), not quite parallel to the original, was determined; on subsequent heating, this new line was followed to nearly 500°C., where a leak developed in the jacket. Surprisingly, the North Carolina dunite showed almost exactly the same behavior, except that the sudden drop took place between 200 and 300. The coefficients found for the "reversible"

TABLE 4.—*Dunite, New Zealand (2+)*

Experimental points at 8500 kg/cm², in order of measurement.

t°C	Condenser setting		Time
23	1506	July 7,	11:45 AM
82	1530		1:50 PM
172	1566		2:35
248	1595		3:00
306	1634		3:09
313	1680		3:30
242	1639		4:00
189	1611		4:25
108	1571		4:48
30	1529		8:13
	overnight at 8500 kg/cm², 30°		
26	1527	July 8,	8:25 AM
82	1556		8:55
163	1598		9:16
168	1600		9:30
251	1644		9:48
257	1646		10:00
320	1687		10:15
325	1687		10:30
373	1715		11:21
381	1716		11:55
441	1781		12:03
453	1782		1:30 PM
548	2150 ?	Jacket	2:00
423	2065 ?	leak	2:20

Calibration: 1783.5 = 11,500 cycles/sec.
1629 = 12,000 cycles/sec.

parts of the curves are given in the tables. Of course, it is not possible to say that heating to higher temperatures would occasion no further "irreversible" changes. These changes are similar in kind, although not in magnitude, to the reductions of frequency on heating found by Ide (1937) and attributed by him to the loosening of the texture resulting from the unequal thermal expansion of the individual crystals. Perhaps this explanation also applies to these cases, but it seems likely that other effects may be involved, such as the annealing of crystals

originally strained. The olivine crystal is somewhat less anisotropic, with respect to thermal expansion, than most of the other important rock minerals, but its volume expansion is greater than that of any except quartz, and it probably retains its hardness to higher temperatures. In

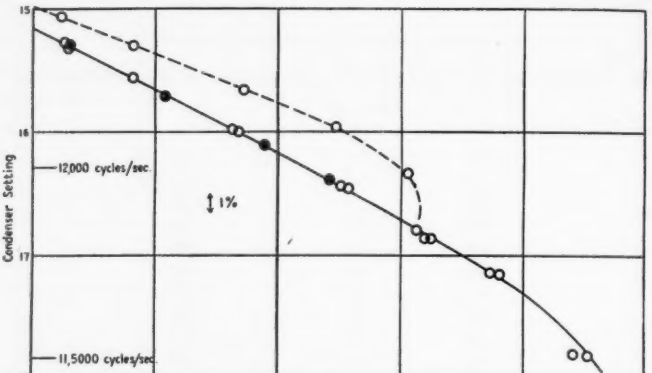


FIGURE 2.—*Original readings for New Zealand dunite at 8500 kg/cm*^{2}

(Condenser settings for resonance as function of temperature.) Open circles denote observations with increasing temperatures; solid circles, observations with decreasing temperatures.

this connection, the effect of the irreversible changes in the quartzite was to *raise* the frequency by about 2 per cent. In most cases, however, the irreversible changes have more effect upon the absolute frequency than on the temperature coefficient.

In a few examples the reverse is true, as, for instance, in the Rockport granite. Here the rate of decrease of frequency as the temperature rose was at first about as low as in the Barre granite; at about 300°C., the rate of decrease began to rise, and on subsequent heating and cooling the initial rate remained about twice as great as its original value, although the absolute value of the frequency had increased only about 1 per cent. The Chelmsford granite started off with about the same rate of decrease as was shown by the Rockport granite after its irreversible change had occurred. The significance of these effects has not yet been satisfactorily worked out.

There are relatively few criteria for testing the validity of these results as temperature coefficients for "perfectly compact" aggregates. The only important rock-forming mineral whose elasticity has been studied at high temperatures is quartz, and even for quartz the data are not complete. Furthermore, the variation with temperature of the elastic con-

stants of quartz at 1 atmosphere is much affected by the position of the transition temperature at 573°C. At 3000 kg/cm², this temperature is displaced to about 640° (Gibson, 1928). However, the data of Atanasoff and Hart (1941) and of Perrier and De Mandrot (1923) show that the effect observed for the quartzite is by no means excessive. For this purpose, we may use the aggregate theory of Voigt, which should give fairly good relative values, if not absolute ones. With the values given for 35° by Atanasoff and Hart, the computed value of V_S for an isotropic quartz aggregate is 4.23 km/sec; at the higher temperatures, not all the constants are given by these authors. The variation of s_{33} is given, however, by Perrier and De Mandrot, and their temperature variation can be applied to the c_{33} of Atanasoff and Hart to obtain a set of values at 500°C. We have also to estimate the values of c_{12} and c_{13} at 500°, in terms of their values at 35° and of temperature coefficients valid in the neighborhood of 35°. Taking $c_{33} = 91.2 \times 10^{10}$, $c_{12} = -13.8 \times 10^{10}$ and $c_{13} = c_{23} = 10 \times 10^{10}$ at 500°, we find at this temperature, $V_S = 4.10$ km/sec. The ratio $4.10/4.23 = .9692$, whereas the observed values of this ratio for the quartzite are 0.9915 at 4000 kg/cm² and .9896 at 3000 kg/cm². The observed changes are considerably less than the computed ones, which depend, unfortunately, upon certain estimates. Another set of figures, less directly comparable, can be obtained from the measurements of Young's modulus for the two principal directions in quartz. The ratio of the value at 400° to the value at 0° is 0.9066 for $E\|$, 0.9588 for $E\perp$; making a correction of 1.019 for the change of density, and taking the square root, we find for the corresponding ratios of "bar velocity," 0.961 parallel to the axis, and 0.988 perpendicular to the axis—again by no means incompatible with the observed values for the aggregate. Thus the indications are that the observed values of the temperature coefficient for the aggregates are probably not much too high as a result of lack of perfect compactness, anisotropy of the crystals, or other possible imperfections. We may also notice that Voigt's theory gives a value 4.23 for V_S of the perfect aggregate, as against an observed value for the quartzite at 4000 kg/cm² of 4.05 km/sec., a difference of about 4 per cent. This difference is probably not greater than can be attributed to the approximate character of the theory itself. Differences as large or larger have been found for metal aggregates (Bruggeman, 1930).

Another indication that the observed temperature coefficient at high pressure is probably close to that for perfectly compact aggregates is its relative independence of the pressure, above perhaps 3000 or 4000 kg/cm², except for quartzite, where the temperature of the $\alpha - \beta$ transition in-

creases by about 20° per 1000 kg/cm². More evidence to support this statement is desirable, but good data for bronzitite at 4000 and at 8000 kg/cm² show closely parallel curves. Even where there are abrupt changes of velocity, the rate of change of velocity with temperature

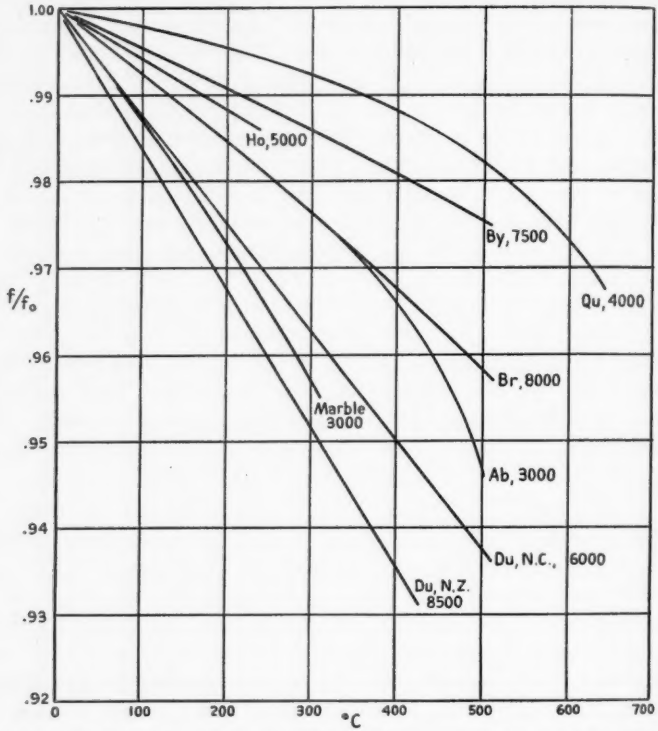


FIGURE 3.—*Relative frequency as function of temperature for various rocks at high pressure*

The numbers attached to curves are pressures, in kg/cm²; f_0 is resonant frequency at 0°C, f at the temperature, t° C. The following abbreviations are used: Qu, quartzite; By, bytownite; Ho, hornblendite; Br, bronzitite; Ab, albitite; Du, dunite.

usually tends to remain nearly constant for the different segments of the curve, but some exceptions to this rule have been noticed.

The principal results are collected in Table 5 and plotted in Figures 3 and 4. In the figures are given the ratio of resonant frequency at high temperature to resonant frequency at 0° C. (f/f_0) at even temperatures, at various constant pressures. In Table 5, the frequency ratios have been

corrected for thermal expansion to give velocity ratios (V/V_0). Except for the quartzite, this correction is minor, so that the lack of good values for the mean thermal expansion of the rocks at high pressure does not introduce appreciable error. For the quartzite, the mean expansion for the quartz crystal at 1 atmosphere was taken; a perceptible error may

TABLE 5.—*Relative velocity at high temperatures and high pressures*

V_0 is the velocity of shear waves at the pressure P and 0° C.

Rock	P Kg/cm²	V_0 Km/sec	V/V₀					
			100° C	200° C	300° C	400° C	500° C	600° C
Quartzite	3000	4.031	.9975	.9948	.9917	.9872	.9806	.....
	4000	4.045	.9980	.9956	.9925	.9883	.9825	.9734
Albitite	3000	3.496	.9922	.9847	.9766	.9661	.9467	.....
Anorthosite	7500	3.712	.9955	.9905	.9859	.9808	.9756	.....
Bronzitite	8000	4.58	.9923	.9846	.9764	.9677	.9584	.....
Amphibolite	5000	4.224	.9942	.9884	.9826	.....	.....	.....
Dunite, N. Z.	8500	4.406	.9833	.9676	.9519	.9358	.....	.....
Dunite, N. C.	6000	4.396	.9872	.9745	.9625	.9509	.9356	.....
Marble	3000	3.483	.9871	.9724	.9572	.....	.....	.....
Granite, Barre	3500	3.59	.9957	.9912	.9870	.9820	.9764	.9625
Granite, Chelmsford	5000	3.596	.9946	.9870	.9755	.....	.....	.....
Granite, Rockport	4000	3.48	.9936	.9858	.9766	.9605	.9371	.9047
Quartz diorite	3000	3.59	.9936	.9874	.....	.....	.....	.....
Diabase, Maryland	3000	3.85	.9895	.....	.....	.....	.....	.....
Gabbro, Mellen	5000	3.758	.9892	.9770	.9623	.9452	.9252	.....
Hornblende—								
garnet	5000	4.32	.9949	.9878	.....	.....	.....	.....
granulite	6000	4.348	.9950	.9814	.9623	.9429	.9238	.....
Eclogite	5000	4.585	.9942	.9884	.9826	.9768	.9710	.9652

result at the higher temperatures because of the displacement of the transition point by the pressure.

A number of the rocks had previously been studied over the range 30° to 100°C. or 30° to 150°C. at 4000 kg/cm². The values of $\frac{1}{V}\frac{(dV)}{(dT)_p}$ given in the older work are compared in Table 6 with the new values for the range 0° to 100°C. The agreement is reasonably satisfactory. The greatest divergences are shown for the Maryland diabase and for the

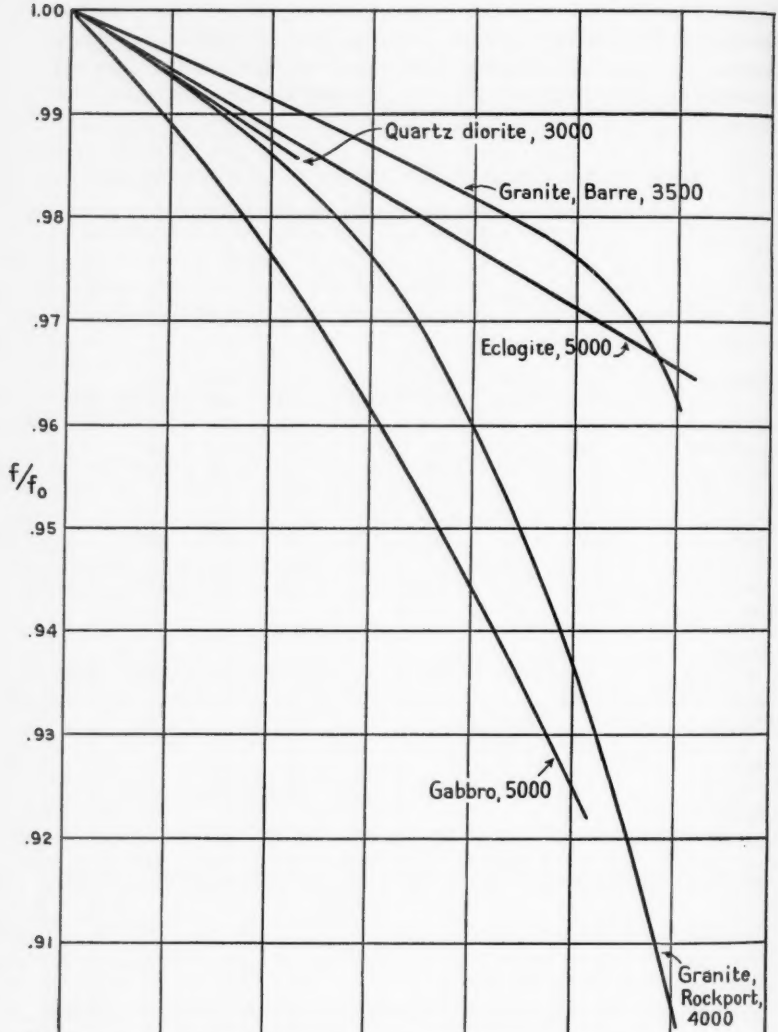


FIGURE 4.—*Relative frequency as function of temperature for various rocks at high pressure*

Symbols have the same significance as in Figure 3.

Rockport granite, already mentioned, where the temperature coefficient on first heating, however, is close to the former one. The coefficients for the dunites are not so high as the highest one found before for the North Carolina dunite, but they remain the highest for the rocks which have been studied.

It is hoped that a thorough discussion based on more extensive material may be possible later; meanwhile the major new results are pointed out and their significance briefly suggested.

The principal result is the extension of the range of temperature available for the determination of velocity at high pressure to about 600°C. Since few authors have suggested average temperatures in the upper 35 km higher than 700°C., the necessity for any substantial extrapolation, with respect to either temperature or pressure, in dealing with the velocities in this region is eliminated. However, data are insufficient to resolve some of the questions about the behavior of particular rocks which are suggested by the peculiarities of the present experimental results. It is desirable to discover why two rocks of nearly identical composition, such as the two dunites, show variations of velocity with temperature differing by 30 per cent and why the velocity in the Barre granite falls off at a rate which is scarcely half that of the Rockport and Chelmsford granites. It remains to be seen whether the application of higher pressures will regularize these effects or introduce even more significant changes in the temperature coefficients.

On the whole, the new results agree surprisingly well with the measurements obtained with a much lower temperature range, as is demonstrated in Table 6. The order of the various minerals and rocks, with respect to temperature variation, remains unchanged to at least 500° C.: for the minerals, the order of increasing temperature effect is quartz, anorthite, amphibole, pyroxene and albite, calcite, olivine; for the rocks, granite, eclogite, diorite, gabbro, marble, dunite. These measurements indicate that the rock giving the highest velocity at temperatures of around 700° C. will be eclogite, rather than dunite. The low coefficient of the eclogite is somewhat surprising and requires further study. A massive garnet aggregate would also be an interesting specimen; one might expect that an aggregate of cubic crystals would be free from the disrupting effect produced by the expansion of noncubic crystals. It is of interest to notice, as shown in Table 6, that except for the position of quartz and calcite the order of increasing (negative) temperature coefficients of velocity is nearly the order of increasing coefficients of thermal expansion; the latter increase in the order, anorthite, calcite, augite and albite, hornblende and diopside, olivine, quartz. In quartz we find a small effect on velocity and a large effect on volume; in calcite, just the reverse.

TABLE 6.—*Effect of temperature upon velocity and volume*

Rock	$-\frac{1}{V_s}\frac{dV_s}{dT}$ at high pressures		Volume thermal expansion at 1 atmosphere (for single crystals) Mean values, 20–400 C
	New values	Former values	
Quartzite	20×10^{-6} 25	9×10^{-6}	49×10^{-6} (quartz)
Albitite	78	77 85	22 (albite)
Anorthosite	45	35 45	15 (anorthite)
Bronzitite	77	76 62 84	25 (diopside)
Amphibolite	58		25 (hornblende)
Dunite	167 128	127 149 208	29 (olivine)
Marble	129	127	16 (calcite)
Granite			
Barre	43		
Chelmsford	54		
Rockport	64	37	
Quartz diorite	64		
Diabase	105	49	
Gabbro	108	116	
Hornblende—garnet granulite	51 50		
Eclogite	58		

The dependence of velocity in a rock upon the velocities in the various mineral components may be obtained very roughly in the following simple way. Let the various minerals, designated as 1, 2 . . .i exist in the pro-

portions, by volume, $x_1, x_2, \ldots x_i$, and let the velocity of a given kind of wave in an isotropic aggregate of each kind of mineral be $V_1, V_2, \ldots V_i$. Then if we suppose that over any path L reasonably long with respect to the mean crystal diameter a ray traverses mineral i in proportion to its amount, by volume, in the rock, we have for the "travel time,"

$$t = \frac{Lx_1}{V_1} + \frac{Lx_2}{V_2} + \ldots\ldots + \frac{Lx_i}{V_i}$$

and the mean velocity V, $V = L/t = 1/\Sigma_i \frac{x_i}{V_i}$

If for the V_i we use the velocities found at high pressure for aggregates of single kinds of minerals, we find that this expression gives a reasonably satisfactory account of the velocities in "mixed" rocks, at high pressure. In using this formula, accessory minerals for which values are not available may be lumped with other minerals which seem likely to have about the same mean velocities; thus the micas can usually be grouped with the feldspars without serious error. For most of the rocks which have been studied, the values calculated in this way agree with the measured values within a few per cent.

We might expect to be able to calculate temperature coefficients of velocity by differentiating this expression with respect to temperature, obtaining:

$$\frac{1}{V}\left(\frac{dV}{dT}\right) = \alpha + V \sum_i \frac{x_i}{V_i} \left\{ \frac{1}{V_i}\left(\frac{dV_i}{dT}\right) - \alpha_i \right\}$$

or the more symmetrical form:

$$\frac{1}{V}\left[\frac{1}{V}\frac{dV}{dT} - \alpha\right] = \sum_i \frac{x_i}{V_i}\left[\frac{1}{V_i}\frac{dV_i}{dT} - \alpha_i\right]$$

α_i is the coefficient of volume thermal expansion of mineral i and: $\alpha = \Sigma_i\, x_i\, \alpha_i$. On substituting the available values, however, we find rather poor agreement with the experimental figures; sometimes the calculated values are too high, sometimes too low, by factors as large as 2. This is not surprising in view of the irregularities of the experimental results which we have already noticed. The formula always gives the same result for a given composition, and, as we have seen, the experimental results for rocks of closely similar compositions may vary perceptibly. No allowance has been made in this very rough theory for such factors as grain size, texture, or preferred orientation of crystals, which may in some cases be of considerable importance.

The way in which such experiments may be used for interpreting seismological results has been indicated in several places (Birch and Bancroft, 1938; Birch, 1941), and no new efforts at identifying the crustal

layers will be made at the present time. On the whole, the new results reinforce the conclusions which have been reached on the basis of the measurements to 100° or 150°C., of which the chief ones are: (1) that the velocity V_S in dunite at a depth of 35 km is not likely to exceed 4.24 ± .04 km/sec; and (2) that at distances over about 20 km the velocity V_S in "granitic" rock in a topmost layer 15 km thick is not likely to be less than 3.4 km/sec. It may well turn out that lateral as well as vertical variations of composition in the crust will have to be taken into account. The values of velocity obtained seismically for a given "layer" may prove to be locally varying averages for several varieties of rock, much as the velocity for a given rock is a kind of average for the different minerals and even for the different directions within each mineral. A uniform layer of homogeneous rock must after all be a greatly idealized first approximation, of which the best test must be comparison of velocities and structures obtained with comparable instruments and techniques in different regions of not too great size. In the compilation of "world" averages from widely differing values, the information which may ultimately prove to be of greatest value in identifying the upper crust may be scrambled in a meaningless way.

REFERENCES CITED

Adams, L. H. and Gibson, R. E. (1926) *The compressibilities of dunite and of basalt glass and their bearings on the composition of the earth,* Nat. Acad. Sci., Pr., vol. 12, p. 275-283.

———, — ——— (1929) *The elastic properties of certain basic rocks and of their constituent minerals,* Nat. Acad. Sci., Pr., vol. 15, p. 713-724.

———, and Williamson, E. G. (1923) *The compressibility of minerals and rocks at high pressures,* Jour. Franklin Inst., vol. 195, p. 475-529.

Atanasoff, J. V., and Hart, P. J. (1940) *Dynamical determination of the elastic constants and their temperature coefficients for quartz,* Phys. Rev., vol. 59, p. 85-97.

Birch, Francis (1939) *Thermoelectric measurement of high temperatures in pressure apparatus,* Rev. Sci. Inst., vol. 10, p. 137-140.

——— (1941) *The problem of identifying the crustal layers,* Am. Geophys. Union, Tr., p. 552-556.

———, and Bancroft, Dennison (1938) *The effect of pressure on the rigidity of rocks,* Jour. Geol., vol. 46, p. 59-87, 113-141.

———, — ——— (1940) *New measurements of the rigidity of rocks at high pressure,* Jour. Geol., vol. 48, p. 752-766.

———, — Clark, Harry (1940) *Thermal conductivity of rocks and its dependence upon temperature and composition,* Am. Jour. Sci., vol. 238, p. 529-558, 613-635.

———, — Law, R. R. (1935) *The measurement of compressibility at high temperatures and high pressures,* Geol. Soc. Am., Bull., vol. 46, p. 1219-1250.

Bridgman, P. W. (1924) *The thermal conductivity and compressibility of several rocks under high pressure,* Am. Jour. Sci., 5th ser., vol. 7, p. 81-102.

Bridgman, P. W. (1925) *Linear compressibility of fourteen natural crystals,* Am. Jour. Sci., 5th ser., vol. 10, p. 483-498.

—— (1928) *The linear compressibility of thirteen natural crystals,* Am. Jour. Sci., 5th ser., vol. 15, p. 287-296.

Bruggeman, D. A. G. (1930) *Elastizitätskonstanten von Kristallaggregaten,* The Hague.

Dale, T. N. (1923) *Commercial granites of New England,* U. S. Geol. Survey, Bull. 738, p. 309.

Gibson, R. E. (1928) *The influence of pressure on the high-low inversion of quartz,* Jour. Phys. Chem., vol. 32, p. 1197-1205.

Ide, J. M. (1937) *The velocity of sound in rocks and glasses as a function of temperature,* Jour. Geol., vol. 45, p. 689-716.

Pabst, A. (1938) *Minerals of California,* Calif. Dept. Nat. Res., Div. Mines, Geol. Branch, Bull. 113.

Perrier, A., and de Mandrot, R. (1923) *Elasticité et symetrie du quartz aux températures élevées,* Mem. Soc. Vandoise Sci. Nat., vol. 1, p. 333-364.

Ross, C. S., Shannon, E. V., and Gonger, F. A. (1928) Econ. Geol., vol. 23, p. 528-552.

Voigt, W. (1928) *Lehrbuch der Kristallphysik,* Leipzig and Berlin (Teubner), p. 954.

Dunbar Laboratory, Harvard University, Cambridge, Mass.

Manuscript received by the Secretary of the Society, May 9, 1942.

Paper No. 79, published under the auspices of the Committee on Research in Experimental Geology and Geophysics and the Division of Geological Sciences at Harvard University.